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CONTENTS

PAGE

REYNOLDS, D. L., and A. HOLMES. The Superposition of Caledonoid Folds on an Older Fold-System in the Dalradians of Malin Head, Co. Donegal 417

HARKER, R. I. Further Data on the Petrology of the Pelitic Hornfelses of the Carn Chuinneag–Inchbae Region, Ross-shire, with Special Reference to the Status of Almandine 445

WALKER, G. P. L., and J. V. ROSS. A Xenolithic Monchiquite Dyke near Glenfinnan, Inverness-shire 463

OLIVER, R. L. Welded Tuffs in the Borrowdale Volcanic Series, English Lake District, with a Note on Similar Rocks in Wales. . 473

GINDY, A. R. The Plutonic History of the Aswan Area, Egypt . . 484

ALLEN, P. Geology and Geography of the London–North Sea Uplands in Wealden Times 498

STRACHAN, I. The Structure and Development of *Peiragraptus fallax*. gen. et sp. nov. A New Graptolite from the Ordovician of Canada 509

DROOGER, C. W. The Oligocene-Miocene Boundary on both sides of the Atlantic 514

CORRESPONDENCE

The Usk Monchiquite—A "Pipe Intrusion" 519

The Great Glen Fault, Co. Donegal 519

REVIEWS
pp. 520–1

PUBLICATIONS RECEIVED
pp. 522–4

INDEX
pp. 525–8

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GEOLOGICAL MAGAZINE

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Issued 7th December, 1954.

The Superposition of Caledonoid Folds on an Older Fold-System in the Dalradians of Malin Head, Co. Donegal

By DORIS L. REYNOLDS AND ARTHUR HOLMES

(PLATE XI)

ABSTRACT

In the Malin Head promontory an Older fold-system, with thrusting and folding about N.W.-trending axes, was followed by Caledonoid folding about N.E.-trending axes. The limbs of the Older folds, both major and minor, now lie in the sheet-dip of the Caledonoid folds, the dips of the Caledonoid limbs being equivalent to the plunges of the axes of the Older folds. Most of the S-planes are shared in common by the two fold-systems. The only exceptions are those that outline the closures and rising parts of the Older fold-system, and these form zones in which the S-planes strike at right angles to the Caledonoid trend. By reference to the outcrop-forms in the area described, and with the aid of plasticine models, it is demonstrated that the outcrop-forms of folded folds are readily recognizable, and that they provide visible evidence of the approximate axial directions and relative ages of the two fold-systems. By comparative study of outcrop-forms on the models and structures seen in sections cut at right angles to each of the fold-axes, it is found that the structures differ radically from those constructed by projecting the outcrop-forms on planes normal to the axes. The " down-the-plunge " method of viewing a geological map thus provides no clue to the actual three-dimensional structure of folded folds.

INTRODUCTION

A FEW years ago we described some spectacular metasomatic replacements of epidiorite by skarn in the White Cow rocks of the Malin Head promontory, Co. Donegal (Holmes and Reynolds, 1947). With these genuine replacements, however, we erroneously grouped what we later recognized to be a tectonic phenomenon (Plate XI, fig. 2), having originally wrongly interpreted it as relics of a quartzite bed the greater part of which in the section illustrated was thought to have been transformed into mica-schist. When we first saw the rocks of the promontory, twenty years ago, we recognized the effects of folding about Caledonoid axes, but we also recognized

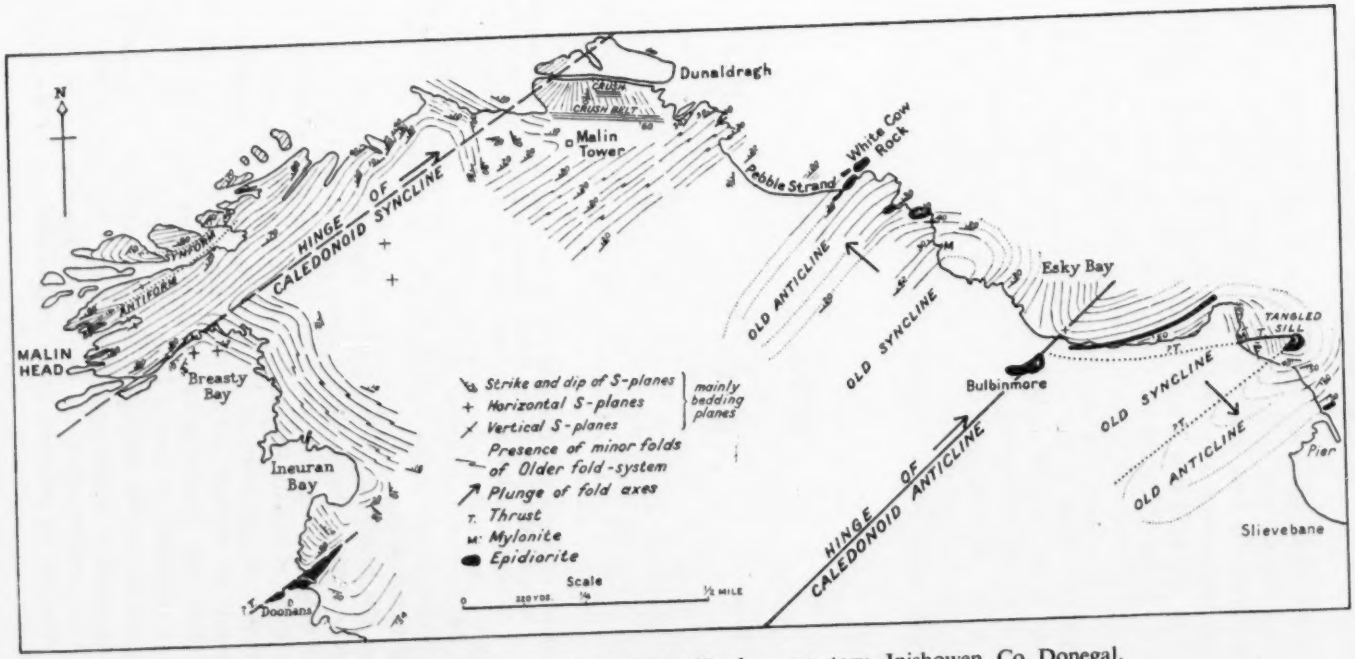
that these effects could not have included the structures here referred to. On returning to the area, in order to investigate the geometry of the fold-system responsible for the appearances that had led us astray, we found that the Dalradian rocks of the area provide convincing evidence that a fold-system with north-easterly trending axes is superposed on a fold-system with north-westerly trending axes. It was lack of knowledge of this older fold-system that led to our mistake.

In the field this superposition of folds suggested to us that the Dalradian rocks of the Malin Head area had been involved in two orogenies, the first being pre-Caledonian in age. Such an inference, however, has far-reaching consequences, and in order not to prejudge the issue—more especially because Dr. B. C. King and others are actively investigating the relationships of corresponding fold-systems in the Scottish Highlands—we have decided to avoid using the term " orogeny ". In order to distinguish the two fold-systems, without any implication as to their interpretation, we shall refer to that with north-easterly trending axes as the " Caledonoid fold-system ", and to that with north-westerly trending axes as the " Older fold-system ".

Our investigation shows that the outcrops of superposed fold-systems form easily recognizable patterns of the sort to which Dr. E. M. Anderson (1951) probably referred when he wrote: " The first simile which occurred to him was that of the teeth of a comb, but the arrangement is perhaps better expressed by the prongs of a fork." The superposition of fold-systems may also account for the remarkable current disagreement between various competent observers as to whether certain lineations are parallel to *a* or to *b*; for obviously, when the axes of two superposed fold-systems are essentially at right angles, the *b* lineation of the first coincides with the *a* direction of the second, and vice versa.

Caledonoid Fold-system

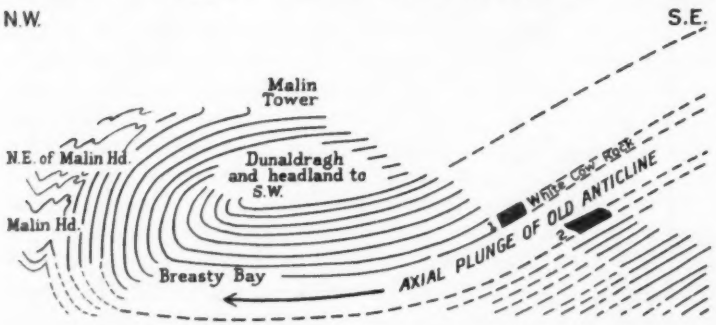
The Malin Head promontory—with an area of just over a square mile—is situated in the extreme north of Inishowen, which lies between Lough Foyle and Lough Swilly. Inishowen is built of metamorphosed sediments and intercalated sheets of epidiorite belonging to the Dalradian, the regional strike of which is north-easterly. The promontory is formed essentially of the Malin Head Quartzite, regarded by McCallien (1935) as the oldest sub-division of the Dalradian of Inishowen. This formation is largely made up of massive to flaggy quartzites with local partings and occasional thicker intercalations of mica-schist that is commonly garnetiferous. Near Dunaldragh (Text-fig. 1), where the schists are particularly strongly developed, they have been conspicuously enriched in albite. Isolated portions



TEXT-FIG. 1.—Structural map of the Malin Head promontory, Inishowen, Co. Donegal.

of sheet-form epidiorite, locally interlayered with the quartzite, provide useful guides in the attempt to unravel the tectonics of what might otherwise have proved to be an intractable area.

The most conspicuous large-scale folds of the area are exposed to view in the bold and precipitous cliffs between Malin Head itself and the island of Dunaldragh (Text-fig. 1). From the top of the cliffs to the west-north-west of Malin Tower, the closure towards the north-west of a recumbent syncline can be seen in the cliffs of the mainland immediately south of the western end of Dunaldragh. From this natural profile and the further exposure behind it, on Dunaldragh itself, it can be seen that the axis of the fold plunges towards the north east at 10°. This plunge can be confirmed by reference to the related small-scale folds of the vicinity. The lower limb of the recumbent syncline, which has a " neutral " posture (Bailey and McCallien,



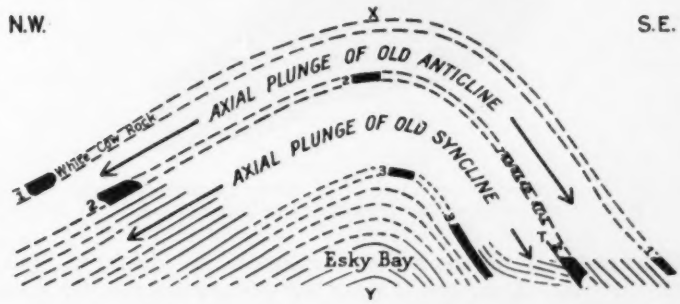
TEXT-FIG. 2.—Composite section of the Caledonoid recumbent syncline between Malin Head and Pebble Strand, constructed from natural sections at various tectonic levels as indicated by the locality names.

1937, fig. 1, p. 81), continues to outcrop along the north-eastern cliffs of the mainland and is demonstrably right way up in Pebble Strand (see p. 431).

Since the hinge-line of the syncline plunges towards the north-east, it is obvious that a slightly lower tectonic level of the hinge should outcrop along the south-west coast of the promontory in the environs of Breasty Bay, to the south-east of Malin Head itself. This closure of the limbs of the syncline is actually spectacularly displayed at the western end of Breasty Bay. Here there is a sudden sharp bend from a steep upper limb, which dips north-westward at 70°, to a practically horizontal lower limb (Text-fig. 2). This sharp bend stands in marked contrast to the gentle curvature of the corresponding synclinal closure near Dunaldragh. The difference in the form of the syncline along the two parallel coast lines is probably to be explained by reference to the relative positions of the major

folds of the Older fold-system. The north-east coast of the promontory is roughly tangential to synclinal closures of the Older fold-system, whereas the south-west coast appears to coincide more nearly with the anticlinal closures of the Older fold-system.

To the west of the closure of the syncline, i.e. to the north of Malin Head itself, the steep north-western limb of the syncline is succeeded westward by the more gently dipping north-western limb of the neighbouring anticline (Text-fig. 2). The recumbence of the anticline towards the south-east, and the forms of the related drag-folds exposed in the north-western cliffs of the promontory, indicate that the direction of movement was towards the south-east. This conclusion is in agreement with Hartley's (1938) observations of the Caledonian folds in the Dalradians of the Sperrin Mountains, Co. Tyrone, and with Bailey and McCallien's (1934) observations of Caledonian folding in the Dalradians of north-east Antrim. In all these areas Caledonoid



TEXT-FIG. 3.—Continuation of Text-fig. 2 from Pebble Strand to the Tangled Sill, showing the Caledonoid anticline of Esky Bay.

anticlines in Dalradian rocks are overturned towards the south-east, like those of the South-Western Highlands of Scotland (Bailey, 1922).

To the south-east of the recumbent syncline, the next anticlineal closure is exposed in plan in the rocky foreshore of Esky Bay. About the middle of Esky Bay there is a change from a north-westerly to a south-easterly dip. The actual hinge of the anticline, here characterized by vertical strain-slip cleavage and by an essentially horizontal disposition of the heavy mineral bands of the quartzite, can be seen to strike north-eastward. The corresponding anticlinal closure on the south-west coast may be expected to outcrop at Binbane, to the south-east of the area described in this paper. Near Doonans, on the south-west coast, the flat-lying quartzite of Breasty and Ineuran Bays and their hinterlands, which form the lower limb of the syncline referred to above, turn sharply upward to form the north-western limb of the anticline under discussion. A thick sheet of epidiorite, located at this

bend, exhibits remarkable crumpling along bands which weather white and appear in sharp contrast against the general dark colour of the epidiorite. In Text-fig. 4 an attempt has been made to depict the very complex small-scale folding of one of these relatively light-coloured bands. The crests of the small folds are overturned towards the south-east, providing additional evidence that the translation was in this direction. Moreover, the hinge-lines of the crests of the small folds plunge north-east at 30°, thus providing additional evidence as to the direction of plunge of the axes of the folds. Two or three feet above the crumpled light-coloured band depicted in Text-fig. 4 a thin sheet of quartzite enclosed within the epidiorite shows recumbent

TEXT-FIG. 4.—Caledonoid crumpling of white-weathering band in the epidiorite of Doonans, south of Ineuran Bay.

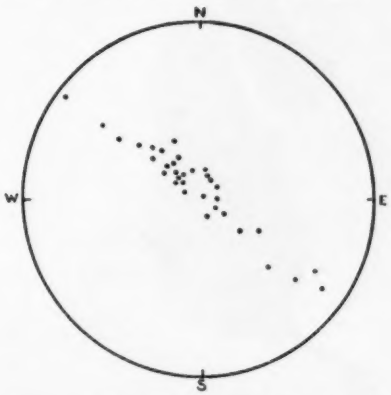
folding related to the more severe crumpling just described. Here again the plunge of the axis is about 30° to the north-east, the movement having been towards the south-east. Evidence as to the mode of emplacement of quartzite layers within epidiorite will be presented in connection with the Older fold-system (p. 425).

The north-easterly plunge of the Caledonoid fold-axes, averaging between 10° and 20°, can be confirmed over and over again by reference to small-scale folds. It must be emphasized, however, that the small-scale Caledonoid folds are locally inconspicuous in comparison with the small-scale folds of the Older fold-system which are present within the same rocks. When the small-scale Caledonoid folds are not related to the closures of the large-scale folds of the same system (e.g. crumpling of schists south-west of Dunaldragh) they are gentle undulations with wave-lengths of 10 to 20 feet or more, and limbs that only rarely dip more steeply than 10°. The dips of the limbs are, of course, towards the north-west and south-east, like those of the larger folds.

In order to confirm the observed gentle angles of plunge and the direction of the Caledonoid fold-axes, the poles of S-planes, mostly bedding planes, which are unambiguously related to the Caledonoid

fold-axes, have been plotted on a stereographic projection (Text-fig. 5). The poles clearly form a girdle extending from north-west to south-east, the position and width of which respectively confirm the north-easterly trend of the fold-axes and their gentle plunge, the latter ranging in general from 0° to 20° to the north-east.

A section across the major Caledonoid folds, essentially at right angles to the fold-axes, is shown in Text-figs. 2 and 3. The section is not a projection of surface outcrops, but a compilation of the natural



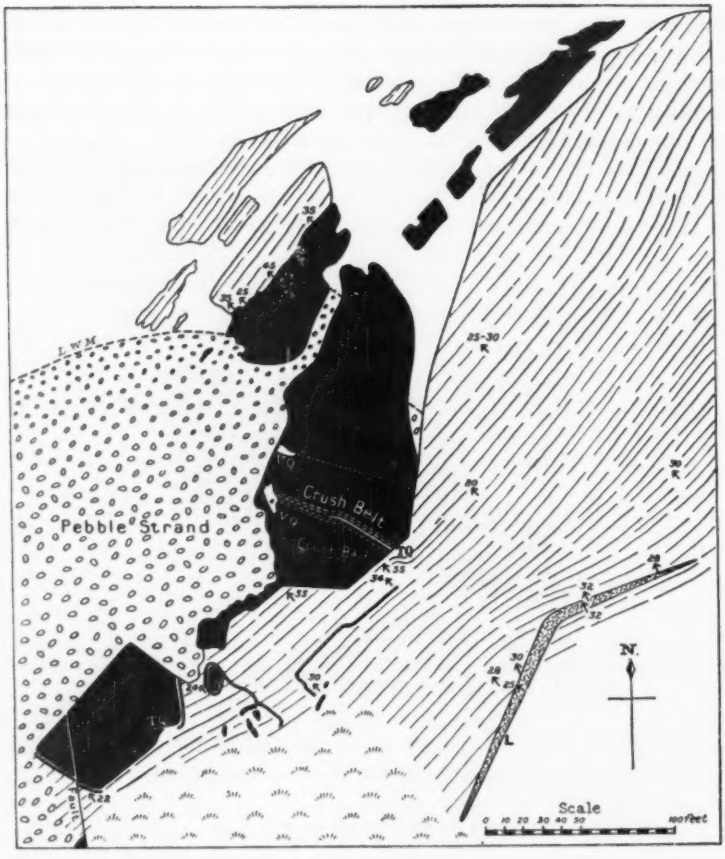
TEXT-FIG. 5.—Stereographic projection (Wulff net) of poles of S-planes (mainly bedding planes) outlining Caledonoid folds.

sections of the different tectonic levels exposed on the south-west and north-east coasts respectively of the Malin Head promontory.

OLDER FOLD-SYSTEM

The Older fold-system is characterized by two phases, an early phase of small-scale thrusting, in imbricate style, followed by a later phase of overfolding towards the south-west about axes striking in a north-westerly direction. The plunge of the axes of these older folds is to the north-west or to the south-east, according to the dip of the limb of the Caledonoid fold structure of which the older fold concerned now forms part.

The first phase is clearly displayed at the eastern end of Pebble Strand. Here a sheet of epidiorite, dissected by marine erosion into a number of isolated segments, forms a headland that is locally known as the White Cow rocks because of the occurrence within the epidiorite of a large mass of white vein-quartz (enclosing an " eye " of amphibole-biotite-skarn) with an outline that simulates the form of a cow's head. A photograph and map of the White Cow rocks have already been published (Holmes and Reynolds, 1947, figs. 2 and 3), and the map,

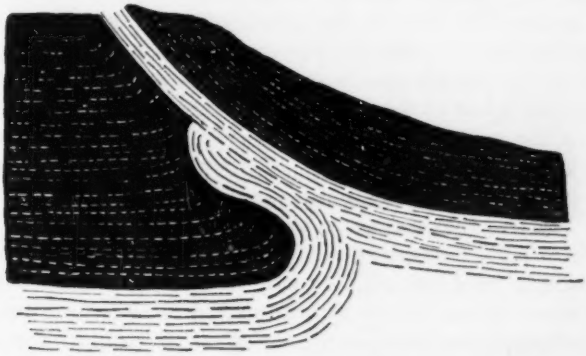


TEXT-FIG. 6.—Map of the White Cow epidiorite (black) and adjoining quartzites and mica-schists (strike indicated by lines), showing the typical mode of occurrence of layers of quartzite (TQ) within the epidiorite. The quartzite layers (TQ) and associated crush-belts mark the positions of thrusts. VQ = vein quartz. L = lamprophyre dyke.

with some modifications, is here reproduced as Text-fig. 6. The sheet of epidiorite is interlayered with the quartzites and subsidiary schists which form part of the north-westerly dipping limb of the Caledonoid anticline whose crest outcrops in Esky Bay. The general strike of the epidiorite and the associated quartzites and schists is north-east, the local dip of 20° to 30° to the north-west steepening towards the epidiorite sheet and reaching 55° at its base. The epidiorite is crossed by crush-belts which are rendered conspicuous by the platy structure of the epidiorite within them. Both the crush-belts themselves and

the lamination of the epidiorite within them strike from 55° to 75° west of north and dip at 60° to east of north. In some of the belts of crushed epidiorite there is a thin median sheet of quartzite, two examples of which are shown on the map, Text-fig. 6. The quartzite within some of these crush-belts is so highly altered as a consequence of metamorphic differentiation, either mechanical or metasomatic, that the occurrence is referred to by Nolan in the Inishowen Memoir (1890, p. 29) as follows: "At Pebble Strand, a large dyke occurs enclosing fragments of gneiss, and so highly altered as to pass in places into hornblende schist." It is our intention to describe this metamorphic differentiation, and other petrological consequences of the tectonic history of the area, in a subsequent paper. Here, however, we are particularly concerned with the tectonic significance of the crush belts and with the process whereby the thin sheets of quartzite came to be emplaced within them.

The most instructive of the sheets of quartzite emplaced within the epidiorite occurs in the crush-belt along which the "cow's head" of white vein-quartz is situated. This sheet of quartzite, which is about



TEXT-FIG. 7.—Elevation illustrating the evidence of emplacement by recumbent-fold thrusting of the layer of quartzite indicated on map, Text-fig. 6, as TQ–VQ, and shown in Fig. 1 of Plate XI.

6 inches thick, is continuous, towards the north-east, with the layer of quartzite immediately beneath the epidiorite sheet (Text-fig. 7, and Plate XI, fig. 1). It is apparent, therefore, that the thin sheet of quartzite forms the sole of an upthrust mass of epidiorite, the upthrusting being towards the south-west. This upthrusting is also evidenced by the fabric of the epidiorite. The epidiorite has a planar foliation parallel to its upper and lower surfaces. The folation of the upthrust mass of epidiorite to the north-east of the thrust is consequently parallel to the thin sheet of quartzite that forms its sole, and thus makes a high

angle with the foliation of the relatively undisturbed epidiorite to the south-west of the thrust (Text-fig. 7). In the most south-westerly mass of epidiorite shown on the map, Text-fig. 6, another example of a thin sheet of quartzite in continuity with the layer of quartzite immediately underlying the epidiorite can be seen. Thrusting of this type is not restricted to the epidiorite. For example, in Pebble Strand, to the south-west of the White Cow rocks, a similar small thrust is exposed in the quartzite-schist series, the thrust plane dipping at 30° towards 21° east of north.

Within the epidiorite there are numerous thin layers of quartzite, displaying various stages of transformation towards quartz-schist (including varieties containing amphibole like that of the epidiorite), for which the exposures do not permit continuity with the underlying quartzite to be seen. Many of these thin sheets of quartzite and its derivatives terminate upwards with an overturn towards the south-west, so that it can be inferred that the thrust planes take the place of the inverted limbs of small-scale folds. That thrusting was prevalent during this early phase of movement implies that the epidiorite and quartzite were at that time too brittle to form unbroken recumbent folds, presumably because the combination of load-pressure and temperature failed to reach the limit necessary for dislocation by rock flowage alone. At the most westerly angle of the south-western mass of epidiorite shown on the map, Text-fig. 6, marine erosion has recently removed the epidiorite from the curved surface of an overturned end of a thin sheet of quartz-schist, and has also, by removal of some of the quartz-schist itself, exposed the underlying epidiorite. Here it can be observed that the plunge of the axis of the overturned quartz-schist layer is 20° towards 54° west of north, the attempted overfolding and thrusting being towards the south-west. Another point of interest in this occurrence is that the exposed curved surface of the epidiorite, which before erosion was in contact with the overlying layer of quartz-schist, is grooved, the grooves striking 38° east of north. This grooving provides an example of lineation in *a* if it is to be referred to the Older folding. It is, however, more probably related to the crumpled quartzite layer in the epidiorite of Ineuran Bay, and similarly dependent on differential movement between epidiorite and quartzite (now quartz-schist) during the later Caledonoid folding. On this interpretation it is lineation in *b*. Incidentally, it may be noted that if only the Caledonoid fold-system had been observed, the thrusting itself would have been mistaken for movement parallel to *b*, the thrust-planes providing *a* lineations. When considered in their proper setting, however, as part of the Older fold-system, the thrust-planes provide evidence of *b* lineations and of movement parallel to *a*.

To the south-east of the White Cow rocks there are small-scale

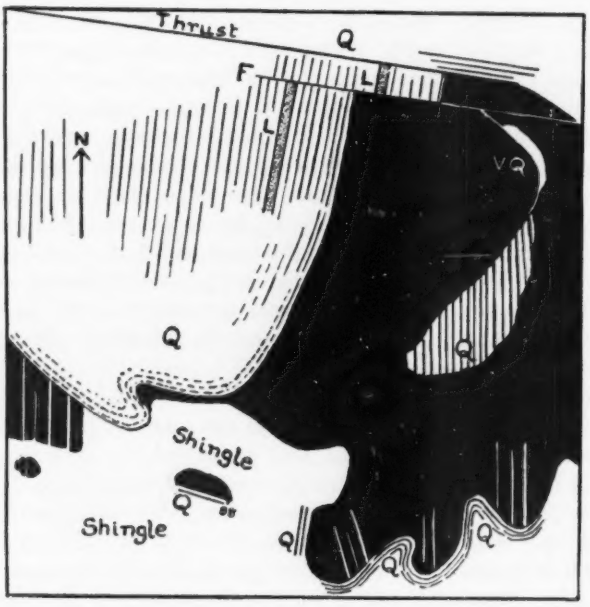
recumbent folds (2 or 3 feet in horizontal extent), overturned towards the south-west. These minor folds are flattened out in the sheet-dip of the Caledonoid structure, to which their axes are parallel, and sometimes they are broken across by closely spaced minor thrusts. On the south-western limb of one such small fold repeated minor thrusts, an inch or less apart, have frayed schist out into quartzite in a south-westerly direction. A larger-scale example of this repeated small-scale thrusting, in rocks immediately underlying the epidiorite, is spectacularly exposed on a steeply inclined rock-face in an embayment to the south-west of the largest outcrop of the White Cow epidiorite. Here the combined effect of the shredding of the rocks by repeated thrusting towards the south-west during the Older fold-period, with the subsequent mangling-out of the resultant tectonic forms into the sheet-dip of the south-western limb of the Caledonoid anticline, has resulted in the appearance illustrated in Plate XI, fig. 2. Not only has the combination of the two fold-systems given rise to " fishes " and boudins of quartzite " swimming " in mica-schist, but the combined effect of shredding and mangling gives, to the unsuspecting petrologist, an illusory appearance of essentially undisturbed bedding. It was this occurrence that we misinterpreted in 1947.

Before leaving the rocks of the White Cow area, mention should be made of the differential movement, along bedding planes, which can be referred to the folding that succeeded the early phase of thrusting discussed above. Thin layers of mica-schist, interbedded with quartzite at this locality, commonly show well developed strain-slip cleavage. That this differential movement dates from the time of the Older fold-system is evidenced by the fact that the cleavage planes dip steeply towards the north-east, thereby indicating that the overlying quartzite in each case has moved south-westward relative to the underlying mica-schist.

Evidence that the early phase of thrusting was indeed followed by one of folding during the Older fold-period is most convincingly displayed by a sheet of epidiorite that outcrops about 300 yards north-west of Slievebane pier. This is probably a disconnected portion of the sheet of epidiorite that outcrops along the eastern end of Esky Bay. It includes numerous layers of quartzite which strike across the width of the epidiorite sheet in a northerly direction, and dip eastward at from 60° to 70° (Text-fig. 8). The thickness of these layers of quartzite ranges from an inch or less up to about a foot, and within the thicker layers the heavy mineral bands are clearly parallel to the boundary planes against the enclosing epidiorite. So complicated is the appearance of this epidiorite sheet with its included layers of quartzite oriented at a high angle to the general strike, that when we first saw it twenty years ago we called it the " tangled sill ". There

is, however, no evidence that it was originally a sill rather than a lava-flow. The appearance of the " tangled sill " is rendered even more complicated by the fact that some of the very narrow layers of quartzite within it show Caledonoid crumpling, like that of the white-weathering bands in the epidiorite of Ineurin Bay.

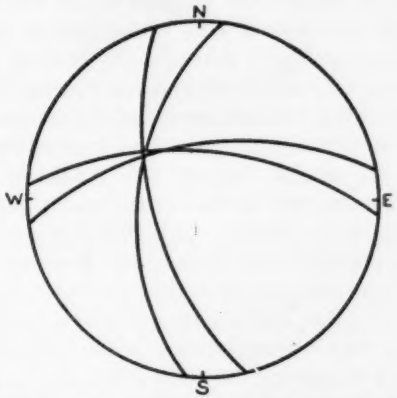
The " tangled sill " provides no evidence as to the method whereby



TEXT-FIG. 8.—Rough sketch-map of the " Tangled Sill " (black), illustrating some of the quartzite layers enclosed within it (white lines) and its drag-folds against the country-rock quartzite (Q: strike indicated by black lines). The spread of quartzite (Q) within the epidiorite is purely diagrammatic. Within the vein quartz (VQ), relics of mylonitized quartzite indicate the continuation of the thrust. L = lamprophyre dyke.

the quartzite layers came to be emplaced within it, but by reference to the White Cow epidiorite it can be inferred that the emplacement was a consequence of repeated thrusting, and that subsequent differential movement between the epidiorite and the country-rock quartzites has obliterated the evidence. There is, however, spectacular evidence that the folding of the epidiorite occurred during the Older fold-period, subsequently to the emplacement of the quartzite layers. On its south-eastern side the epidiorite shows drag-folds, measuring from 8 to 12 feet across, against the neighbouring quartzite (Text-fig. 8).

The plunge of the axes of these drag-folds can be judged to be towards the south-east, but cannot be directly measured. Four reliable measurements of dips and strikes of differently oriented parts of the folded plane of contact between epidiorite and country-rock quartzite were therefore made, and plotted as great circles on a stereographic projection. If the folding that gave rise to the drag-folds were perfectly cylindroidal, then the great circles, defining the contact plane of the drag-folds should intersect in a point, the point being the pole of the fold-axis (Sander's method, as reported by Weiss, 1954). As can be seen from Text-fig. 9, the intersection of the great circles outlines a small triangle of errors. By taking the appropriate point within this triangle to represent the pole of the fold-axis, it appears that the axis plunges at 44° towards 40° south of east. The drag-folds were therefore formed during the Older fold-period. That the folding was subsequent to the early phase of thrusting responsible for the emplacement of the quartzite layers within the epidiorite is evidenced by the fact that these



TEXT-FIG. 9.—Stereographic projection (Wulff net) of four measured S-planes outlining the drag-folds of the Tangled Sill (Text-fig. 8).

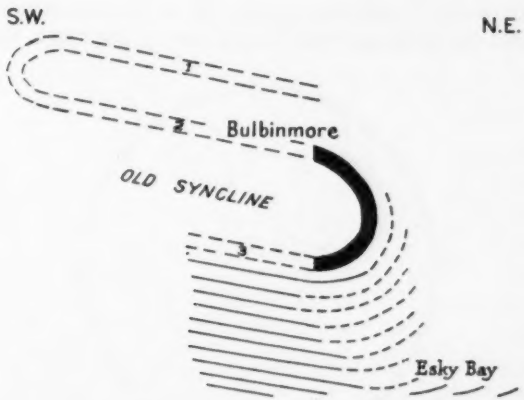
enclosed layers are abruptly truncated by the country-rock quartzite in contact with the drag-folds. Moreover, there is a conspicuous difference between the respective strike directions of the heavy mineral bands of the two sets of quartzite.

The most striking evidence of large-scale folding belonging to the Older fold-system is provided by the rocks of Esky Bay and its environs. To the south-west of Esky Bay, epidiorite, capping the hill of Bulbinmore, forms part of the closure of the Caledonoid anticline, the north-easterly trending hinge-line of which outcrops in Esky Bay. From the top of the hill, standing on the epidiorite, the trend of the hinge-line running out into the Bay is clearly visible, because it coincides

with vertical strain-slip cleavage. To the west of the hinge-line the rocks dip north-westward at 30°, but their strike, instead of remaining parallel to the Caledonoid direction, gradually swings round in an anticlockwise direction, forming an arc that is convex northwards (Text-fig. 1). This arcuate strike curves round the headland at the western end of Esky Bay. To the east of the hinge-line the dip is towards the south-east at about 50° and on this side the strike gradually swings round in a clockwise direction, again forming an arc that is convex northwards. Here the arcuate strike curves around the headland at the eastern end of Esky Bay. These two arcuate outcrops, which have the appearance of two coalescing folds curving away from one another like mirror images reflected across the Caledonoid anticlinal axis, are the surface expression of a major recumbent syncline belonging to the Older fold-system (closing northwards like the minor synclines of that fold-system) which is folded in the Caledonoid anticline. The significance of the mirror-image forms can be appreciated if a tracing is made of the arcuate outcrop of the old syncline to the north-west of the Caledonoid axis, whilst keeping the edge of the tracing paper coincident with this axis. By holding the edge of the tracing paper along the Caledonoid axis and rotating the tracing paper through 180° about this axis, the tracing of the arcuate outcrop of the syncline approximately describes the folded form of the old recumbent syncline. After the rotation through 180° the traced outcrop of the old syncline is seen through the back of the tracing paper and appears as a mirror-image of the synclinal arc in its initial position. When the tracing paper has been rotated through 90° from its initial position, so that it extends at right angles to the surface of the map, it can be appreciated that, with an appropriate topographic relief (and making due allowance for the exaggerating effect of the obliquity of the section of which the arc is a tracing), the limbs of the old recumbent syncline may outcrop for a short distance, with a strike at right angles to the Caledonoid direction, and with a north-easterly dip. The outcrops of epidiorite on Bulbinmore, already referred to as forming part of the closure of the Caledonoid anticline, can be interpreted in this way (Text-fig. 10). Not only does the epidiorite strike at right angles to the Caledonoid trend for a short distance, but it also extends intermittently down the north-easterly face of the hill through a vertical distance of about 150 feet, dipping in part at an angle that is considerably steeper than the plunge of the Caledonoid axis in the same direction. The epidiorite represents the rise of the old syncline from its lower to its upper limb (Text-fig. 10).

To the north-west of the Caledonoid axis of Esky Bay, the existence of an old anticline at a higher structural level than the old syncline can be inferred from the fact that the rocks, although they must inevit-

ably be inverted in the upper limb of the recumbent syncline, are again right way up (as evidenced by the disposition of the thrusts) at the White Cow rocks of Pebble Strand. Lack of exposures makes it impossible to map the anticlinal closure in either of the mirror-image forms. It is significant, however, that the two outcrops of epidiorite to the north-west of the Caledonoid hinge-line are repeated, a similar distance apart, in the mirror-image form to the south-east of the Caledonoid hinge-line. The epidiorite of the White Cow rocks forms part of the upper limb of the recumbent anticline of the Older fold-system, and



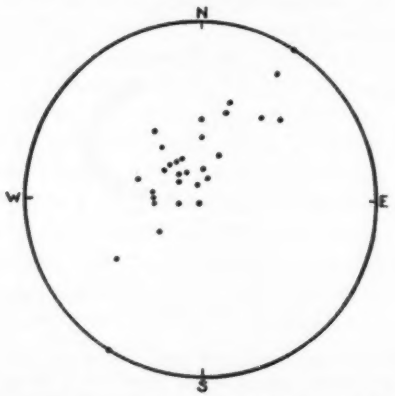
TEXT-FIG. 10.—Diagrammatic section drawn through XY of Text-fig. 3 at right-angles to the plane of the paper, illustrating the way in which the rise of the old syncline results in a spread of epidiorite (black) down the north-eastern face of Bulbinmore, thus providing a connecting link between the occurrences of epidiorite marked 2 and 3 in Text-fig. 3.

it can be inferred that the epidiorite of the " tangled sill " forms part of the lower limb in the mirror-image form. Reference to the map, Text-fig. 1, will show that, although there is no proof, it is quite possible that all the isolated occurrences of epidiorite are portions of a once continuous sheet.

As can be seen from the outcrops of the mirror-image forms, the limbs of the recumbent folds of the Older fold-system are isoclinal, and parallel to the sheet-dips of the limbs of the Caledonoid anticline of which they form part. The fold-axes of the major folds of the Older fold-system therefore inevitably lie in the sheet-dip of the limbs of the Caledonoid anticline, with the fold-axes plunging north-westward where they form the north-western limb of the Caledonoid anticline, and south-eastward where they form its south-eastern limb (Text-fig. 1).

In the small bay to the south-east of Esky Bay the presence of a

thrust, striking eastward towards the " tangled sill " and dipping southward at about 60° is revealed by the presence of mylonized quartzite. The thrust-plane is undulated by Caledonoid drag-folds, the axes of which plunge towards the south-west. The mylonite reaches a maximum thickness of 10 inches in the crests of the drag-folds and thins away to nothing on their limbs. Within the " tangled sill ", mylonized quartzite occurs in a mass of vein-quartz and marks the continuation of the thrust, the strike of which can here be judged to be approximately north-west. From the folding and flowage of the mylonite in the Caledonoid drag-folds, and from the relationship of the thrust to the lower limb of the old anticline, it can be inferred that the mylonite was formed during the Older fold-period.



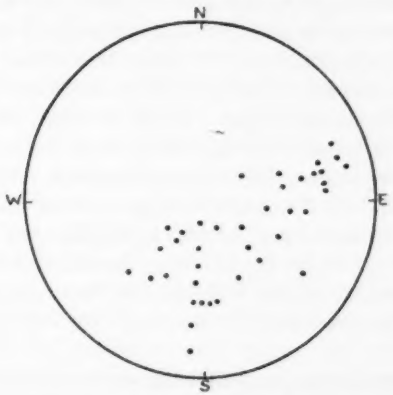
TEXT-FIG. 11.—Stereographic projection (Wulff net) of poles of S-planes (mainly bedding planes) outlining the Older folds located on north-westerly dipping limbs of the Caledonoid folds.

This inference is confirmed by the reappearance of mylonite in the corresponding position within the mirror-image replica of the old anticline on the north-western side of Esky Bay. The reappearance of the mylonite shows it to have been folded by the major Caledonoid anticline.

The above interpretation of the mirror-image outcrops diverging from the middle of Esky Bay can be confirmed by reference to the related small-scale drag-folds described below. It can also be confirmed by plotting on stereographic projections the poles of S-planes (mostly bedding planes) defining the mirror-image forms. The poles of S-planes from the mirror-image form to the north-west of the hinge-line of the Caledonoid anticline are plotted in Text-fig. 11, and those from the mirror-image form to the south-east of this hinge-line are plotted in Text-fig. 12. In each of these figures the poles of S-planes form

girdles about a north-westerly trending fold-axis, but in Text-fig. 11 the plunge is to the north-west, whereas in 12 it is to the south-east, each being in conformity with the respective dip of the limb of the Caledonoid anticline on which the mirror-image form lies. The breadth of the girdles expresses both the variation in the angle of dip of the respective limbs of the Caledonoid fold, and the variation in plunge of the axes of the mirror-image folds. If the Caledonoid fold were smoothed out, then the plunges of the older folds would disappear, and the girdles on the two stereograms (11 and 12) would narrow down until they coincided as a diameter extending from north-east to south-west.

When arcs are drawn on a stereographic projection (Wulff net)



TEXT-FIG. 12.—Stereographic projection (Wulff net) of poles of S-planes (mainly bedding planes) outlining the Older folds located on the south-easterly dipping limb of the Caledonoid anticline of Esky Bay.

to represent the positions in space of S-planes of the promontory, a very high proportion of the arcs pass through, or close to, the fold-axes of both the Caledonoid fold-system and the Older fold-system. This demonstration that a high proportion of the S-planes are essentially parallel to the axes of both fold-systems is an expression of the fact that the limbs of the folds of the Older fold-system lie in the sheet-dip of the Caledonoid fold-system. In other words, with the exception of the S-planes outlining the closures of the Older fold-system, the S-planes of the promontory are parallel to the axes of both fold-systems. The S-planes outlining the closures of the Older fold-system, however, intersect the Caledonoid axes at a high angle, and strike approximately at right angles to the Caledonoid trend. Thus the superposition of folding about a Caledonoid axis on an Older fold-system with north-westerly trending axes, results in the

intersection of the present-day Caledonoid pattern by undulating belts in which the strike is essentially north-west. Apart from the widespread evidence of minor overfolding towards the south-west about north-westerly trending axes, it is in these undulating belts, cross-cutting the Caledonoid trend, that the evidence of major folds of the Older fold-system is to be found.

The small-scale anticlines related to the Older fold-system are overfolded towards the south-west about north-westerly trending axes. When they are located on the north-westerly dipping limb of the Caledonoid anticline their axes plunge north-westward, whereas the axes of the small-scale folds on the south-easterly dipping limbs of the Caledonoid folds plunge south-eastward. In the flat-lying rocks of Breasty and Ineurin Bays, the axes of the small-scale folds of the Older fold-system plunge at gentle angles or not at all.

In the steeply dipping rocks of Malin Head itself and its vicinity one might hope to find small-scale folds of the Older fold-system with almost vertically plunging axes. Here, however, the rocks exhibit closely spaced strain-slip cleavage, and the small-scale folds of the Older fold-system seem to have been obliterated. There is evidence, however, indicated on the map (Text-fig. 1), that the larger folds of the Older fold-system have steeply plunging axes. The limbs of these larger folds can be traced from the blurred relics of colour-banding indicative of original bedding, the blurring appearing to have resulted from the shredding of the rocks by closely spaced minor thrusts.

While there can be no doubt that the small-scale folds of the Older fold-system were initially recumbent, their present-day style, with the lower inverted limb commonly equal in length to the upper limb, probably owes much to the fact that they have been flattened out in the sheet-dip of the Caledonoid structures to which their isoclinal limbs are approximately parallel, and in the plane of which their axes essentially lie. This mangling out of the minor folds of the Older fold-system into the sheet-dip of the younger Caledonoid folds has not only lengthened the recumbent portions of their limbs, and particularly of their lower limbs, but it has probably also reduced the length of the "normal" and "rising" (Bailey and McCallien, 1937, fig. 1, p. 81) portions of the limbs. Differential movement between the S-planes during Caledonoid folding has resulted in minor folds of the Older fold-system becoming sandwiched between layers that have the appearance of being undisturbed, but which may themselves show similar folding only a few feet away. As a consequence of the flattening out of the minor folds of the Older fold-system into the sheet-dip of the major Caledonoid folds, their axes, although at right angles to those of the minor Caledonoid folds, lie essentially in the same plane.

As already described, however, the minor Caledonoid folds have a gently undulatory form.

Plasticine Models

In order to test our interpretation of mirror-image fold-forms as the surface outcrops of folded folds, we made two plasticine models.

For the first model four pounds of plasticine were used—a pound each of four different colours. Each colour was first made into a rectangular sheet, measuring about 12 inches by 8, and the sheets were then placed one on top of another in a recorded stratigraphic sequence of colours. The pile of layers was then folded into three recumbent folds about axes parallel to the shorter edge of the rectangle. The resultant folded mass was then itself folded, about axes at right angles to those of the first folding, into two anticlines with an intervening syncline. The doubly folded model was then placed on a cardboard base on which the directions of the two fold-axes were recorded.

In Text-fig. 13 (*c*) a tracing of a section of the model cut at right angles to the axis of the first folding and along the crest of an anticline of the second folding is shown. This section reveals the form of the folds of the first folding. A tracing (13 (*d*)) of a section of the model cut at right angles to the axis of the second folding shows the form of the folds of the second folding.

The next step was to denude the model by cutting slices from the top in order to discover the appearance of the surface outcrops corresponding to the known internal structure. The first cut, made to produce a flat upper surface, revealed the outcrops of mantled domes (Text-fig. 13 (*b*)), the domes coinciding with the intersection of anticlines of the second folding with anticlines of the first. The outcrops of the mantled domes, as shown in 13 (*b*), are quite disconnected, although the domes themselves have below-surface connections. The outward dips of the domes are accounted for by the dips of the limbs of the anticlines of the second folding, and by the dips of the closures and rising parts of the anticlines of the first folding.

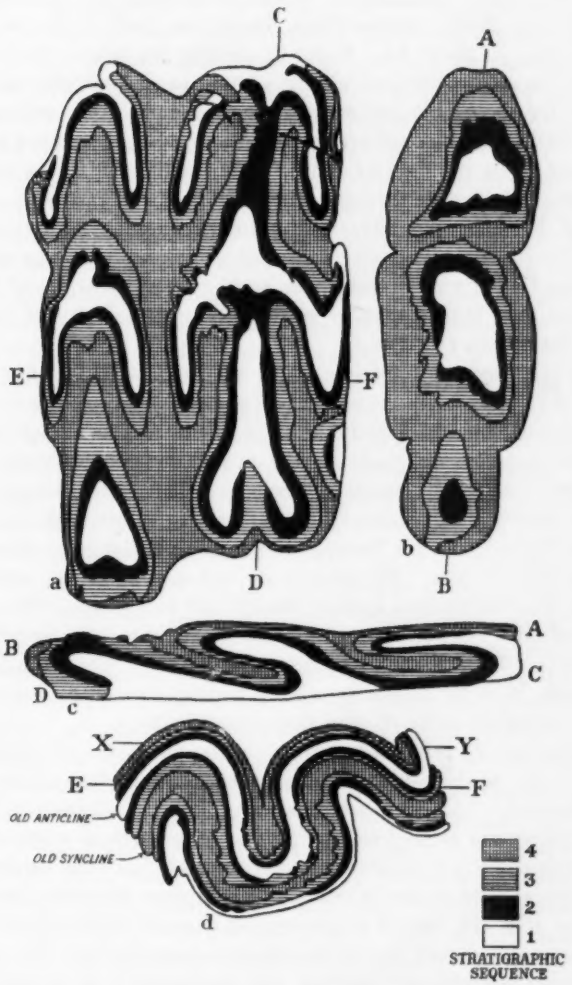
The surface of the model was then denuded down a little further by cutting off another slice. The outcrops now appeared as mirror-image forms, the axes of symmetry of which coincide with the anticlinal axes of the second fold-system. A tracing of the surface of the model as it appears at this level is shown in Text-fig. 13 (*a*), E–F being the direction of the first fold-axis and C–D that of the second. The mirror-image forms on the left-hand side of 13 (*a*) differ from those on the right-hand side because the surface of the model (E–F in 13 (*d*)) cuts across the anticlines of the second folding at a higher level on the left as compared with the right. In other words, the mirror-image forms on the left of 13 (*a*) represent a tectonic level intermediate between

that of the mantled dome outcrops (13 (*b*)) and that of the mirror-image forms on the right of 13 (*a*).

As was discovered from the rocks of the Malin Head promontory, it is apparent by comparing (*a*), (*c*), and (*d*) of Text-fig. 13 that the structural planes in the plasticine model are common to both fold-systems and parallel to both fold-axes, except for those outlining the closures and rising portions of the first fold-system. These latter give rise to outcrops striking in general at right angles to the trend of the axes of the second fold-system (Text-fig. 13 (*a*)) and having steep or vertical dips. In detail, however, the outcrops have an arcuate form, since the outcrops of the closures and rising parts of the first folds gradually curve round to the limbs of the first folds that lie in the sheet-dip of the second folding.

By reference to the outcrops of mirror-image forms on the plasticine model (Text-fig. 13 (*a*) and (*b*)) it is possible to recognize superposed fold-systems in which the axes are approximately at right angles to one another, from the form of the outcrops on geological maps. It is, moreover, possible to find the direction of strike of the axes of the two fold-systems from the surface outcrops. By comparing (*a*), (*c*), and (*d*) of Text-fig. 13, it can be seen that the direction of the axis of the first folding can be found by joining the points of the " prongs " (outcropping at the same topographic level) in the mirror-image fold-forms, i.e. by joining the anticlinal closures in mirror-image forms, or by joining the related synclinal closures. The direction of the axis of the second folding coincides with that of the axis of symmetry of the mirror-image forms.

Characteristic examples of mirror-image forms can be seen on many of the one-inch geological maps of the Scottish Highlands. On Sheet 46, for example, a distorted mirror-image form appears in the extreme north-east corner of the map. Disregarding the distortion, the symmetry axis of this form indicates that the axis of the second folding has a Caledonoid trend, at a high angle to that of the first folding, the exact direction of which cannot be determined in this instance by joining the " prongs ", because the latter are exposed at very different topographic levels. However, the axial direction, varying from west of north to north-west, can be read over and over again on this map by joining the tips of " prongs " in much smaller mirror-image forms (" the teeth of a comb ") in the Loch Tay Limestone, e.g. in the north-east corner of Sheet 46 and on the western side of the map to the south-west of the River Fillan. Another example of a mirror-image form on Sheet 46 is provided by the epidiorite in the north-west of the map. Here the axis of the first folding, found by joining the prongs, strikes slightly west of north, whilst that of the second (i.e. the symmetry axis of the mirror-image form) strikes slightly



TEXT-FIG. 13.—First plasticine model of folded folds. EF is the axial direction of the first folding, and CD (or AB) that of the second.

(*a*) Surface outcrops after slicing.

(*b*) Surface outcrops on the right-hand side only, as seen at a higher level than that of (*a*).

(*c*) Section of the model, before slicing off its top, cut along CD of (*a*) and AB of (*b*). AB and CD represent the levels at which (*b*) and (*a*) were respectively sliced.

(*d*) Section of the model, before slicing off its top, cut along EF of (*a*). XY and EF represent the levels at which (*b*) and (*a*) were respectively sliced.

north of east. Considerably farther north-east, on the borders of Moray and Banff, mirror-image forms are well displayed on the eastern side of Sheet 85. Here, by joining the tips of the quartzite prongs, the axis of the first folding is found to strike north-west, whilst that of the second, by reference to the symmetry axis, strikes north-east. Further examples of prongs and mirror-image forms are provided on Sheet 85 by the lines of contact between the " Black Schist " and the Slate and Phyllite Group.

Dr. E. M. Anderson (1951) undoubtedly pointed to a most significant clue towards unveiling the secret of Highland tectonics when he recognized what may be called *the problem of the prongs* and directed attention to it by speaking so graphically of " the teeth of a comb " and " the prongs of a fork ".

Two other lessons are to be learnt from the plasticine model. The first is that when surface outcrops represent folded folds, a projection of these outcrops on a plane at right angles to the plunge of either fold-axis supplies no guide to the actual geometrical form of the structure. This can be readily appreciated by comparing the real section at right angles to the second fold-axis (Text-fig. 13 (*d*)) with the appearance of the outcrops on Text-fig. 13 (*a*), as seen along the direction of the second fold-axis. The prongs do not appear in the section. It follows that the method, used by Bailey and McCallien (1937) in their study of the " Schiehallion twist " and subsequently amplified by Mackin (1950), of holding a map so as to look at it down the direction of plunge, in order to obtain a good general idea of a cross-section, is not applicable to maps depicting folded folds.

In illustration, reference may be made to J. G. C. Anderson's recent work in the Slieve League promontory in the south-west of Co. Donegal. The map accompanying J. G. C. Anderson's paper (1954) shows the axes of minor folds to strike both in north-westerly and north-easterly directions. Moreover, the fact that vertical beds also strike at right angles to one another, in these same two directions, indicates that the major structures also result from folding about north-westerly as well as north-easterly trending axes. By applying the " down the plunge " method and reproducing, as a section, the prong-like forms of outcrops seen on the map, J. G. C. Anderson has drawn a mechanically impossible and non-existent section, by reference to which he attempts to explain thrusting in opposite senses. Nevertheless, J. G. C. Anderson's observations of the directions of strike of axes of minor folds are of great importance, since they indicate that the two fold-systems of the Malin Head promontory in the extreme north-east of Co. Donegal are also present in the extreme south-west of Co. Donegal. That folded folds are also present in the intervening area is apparent from the two main directions of lineation,

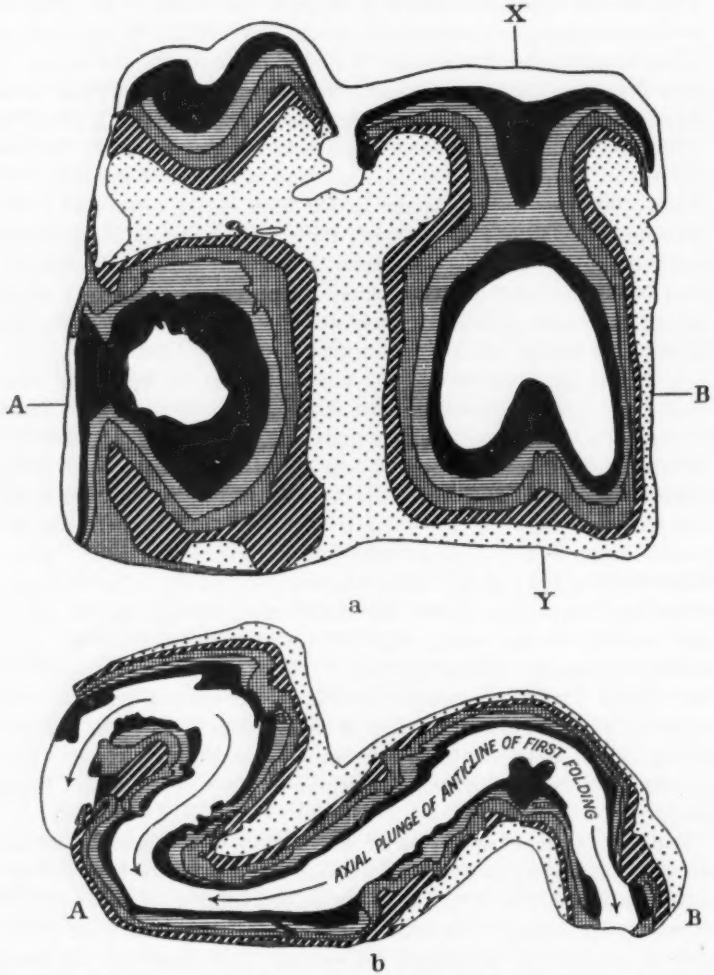
both seemingly *b*-lineations, trending respectively north-north-west and west-south-west, and also from the strike of vertical beds in these same two directions (Pitcher, 1953). It is worthy of note that Pitcher records his opinion (1953, p. 444), with reference to these structures, that "a hypothesis of two separate movement phases might well resolve some of the complex evidence".

The second lesson to be learnt from the model relates to the difficulty involved in attempts to unravel the structure of folded folds by reference to the stratigraphic succession. Sections (*c*) and (*d*) in Text-fig. 13 show that after the second folding the beds are repeated three times in a high proportion of possible sections. This leads to a doubling of the thickness of the oldest and youngest strata affected by the two foldings in the cores of the anticlines and synclines of the first fold-system. Another important fact is that both on the limbs and in the closures of the second fold-system, the strata appear in both correct and inverted stratigraphic succession. Uncritical reliance on observations of current- and graded-bedding thus becomes practically valueless, and may even be positively misleading, in deciphering the structure by reference to the stratigraphy in regions of folded folds.

A second plasticine model was made in order to reproduce more exactly the disposition of the structural planes in the Malin Head promontory. For this eight pounds of plasticine of six different colours were used, each pound being made into a rectangular sheet, about 12 inches by 8, as before. This time, however, in piling the sheets in a recorded sequence, two pounds of one colour were used for the bottom layer, and two of another colour for the top layer. Thus, before folding, the top and bottom layers each had twice the thickness of the four intervening layers. The first folding was carried out about axes parallel to the longer edge of the rectangular pile, the form produced being as shown in Text-fig. 15 (*b*). The folded mass was then itself folded into a recumbent anticline, a recumbent syncline, and a normal anticline, the axes being at right angles to those of the first folding. The model was then oriented on a board, with its axes and fold-forms in the same relative positions as those of the Malin Head promontory.

Text-fig. 14 (*b*) is a tracing of the section seen when the model was cut at right angles to the direction of the axes of the second fold-system and illustrates the form of the second folding. Text-fig. 15 (*a*) shows the section seen when the model was cut along the crest of the normal anticline of the second fold-system, and illustrates the form of the folds of the first system as they appear after the second folding. The upper part of the model was sliced at appropriate levels to reproduce as nearly as possible the surface outcrops of the Malin Head promontory. Although no attempt was made to reproduce the detail

of the topography, the normal anticline was cut at a lower structural level than the recumbent anticline, and sloping slices, approximately parallel to the axes of the first folding, were cut from the recumbent anticline in order to reveal the structure corresponding to that exposed in the cliff sections of Malin Head. Text-fig. 14 (*a*), a tracing of the surface of the model thus completed, reproduces the dome structure of

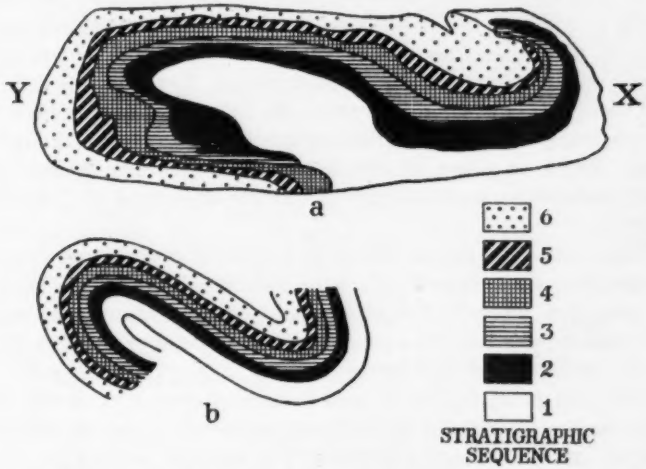


TEXT-FIG. 14.—Second plasticine model of folded folds. AB is the axial direction of the first folding and XY that of the second. For the sequence of layers see Text-fig. 15.

(*a*) Surface outcrops after slicing.

(*b*) Section of model, before slicing off its top, along AB of (*a*).

the western part of the promontory, the flat-lying beds of Breasty and Ineuran Bays, the arc-like mirror-image forms of Esky Bay, the steeply dipping layers striking north-westward to the south-west of Esky Bay, and the divergent strikes exposed on the headland to the south of Dunaldragh. Text-fig. 14 (*b*) reproduces the section across the Caledonoid fold structures of the promontory very closely (cf. Text-figs. 2 and 3), even down to such details as the difference of form shown by the higher and lower tectonic levels exposed along the north-east and south-west coasts respectively. The gently curving closure in the core of the recumbent syncline, for example, is represented by a sharp bend (layer 2) at a lower structural level in the model, as in Breasty Bay. Moreover, the sharp upward bend of layer 4 on the



TEXT-FIG. 15.—(*a*) Section of the second model, before slicing off its top, cut along XY of (*a*) in Text-fig. 14.
(*b*) Section of the second model, after the first folding and before the second, cut at right-angles to the fold-axis.

lower limb of the recumbent syncline in the model resembles the sharp upward bend of the quartzite and epidiorite at Doonans, south of Ineuran Bay. The upward extension, as a heart-shaped form, of layer 2 in the closure of the normal anticline of the model represents the closure of the anticline of the first folding (Text-fig. 15 (*a*)), and is a consequence of the coincidence of the line of section of Text-fig. 14 (*b*) with the anticlinal closure of layer 2 in Text-fig. 15 (*a*). The extension of the epidiorite up the north-eastern slopes of Bulbinmore, referred to on page 430, similarly corresponds to the rise of the lower limb of a syncline of the first folding towards an anticline at a higher structural level.

The second model, like the first, brings out some remarkable relationships between stratigraphy and structure which may well be helpful—both as guide and as warning—in attempts to unravel the intricacies produced by superimposed folding. For example, it can be seen from Text-fig. 14 (*b*) that not only the core of the syncline, but also that of the normal anticline, is occupied by layer 6, the " youngest " of the plasticine sheets. It is thus made apparent that if two fold-systems are involved, even the method of distinguishing anticlines from synclines by reference to the relative age of the bed occupying the core of the fold is unreliable.

In order to expose the structural pattern to be seen at a lower level than that illustrated in Text-fig. 14 (*a*), the recumbent fold was cut so that another slice could be lifted from the left-hand side of the model at will. At this lower level mirror-image forms appear which differ in plan from those portrayed on the right-hand side of 14 (*a*) only in having more widely spaced-out prongs. But in the third dimension there is a significant difference—the forms in the recumbent fold dip isoclinally to the left. In this respect they resemble certain mirror-image forms on Sheet 46 (Scotland) which are also isoclinal and in all probability constitute part of a larger recumbent fold, as in the model.

Finally, attention may be drawn to a point of considerable interest: namely, that the structural results of differential movements between the plasticine layers correspond much more faithfully than might have been expected to those actually observed between the rock layers of the Malin Head promontory. This correspondence serves to indicate that plasticine has a combination of properties of the right order, on the scale of the models here described, to match physically those of the Malin Head quartzites. For example, a model in which length is reduced from L in the field to l in the model, and density similarly from D to d, should be made of material having a strength s which is related to the strength S of quartzite by the expression:

$$\frac{s}{S} = \frac{l}{L} \times \frac{d}{D}$$ (cf. Hubbert, 1945, p. 1640)

Comparison between map and model shows that $L/l = ca.$ 16,500. For the average plasticine used, $d = 1{\cdot}85$. The Malin Head quartzites average about $D = 2{\cdot}65$. Hence, we have $s = S/23{,}600$. The value to be assigned to S is not easy to determine from the available data. An average value for the strength of quartzite is about 30,000 lb./sq. in., the maximum going up to 45,000. The Malin Head quartzites are now all highly recrystallized and have probably well over average strength. But on the other hand the effective strength during the process of folding and recrystallization may have been less than the present

maximum. However, adopting the average, for want of a better guide, the strength of an appropriate material for the model would be about 30,000/23,600 = *ca.* 1·3 lb./sq. in. A few rough tests showed that the average strength of the plasticine, as used, was about 1·8 lb./sq. in., which is clearly of the order required by the principle of physical similarity. The actual strength of plasticine varies somewhat with its colour (i.e. composition), and also with temperature and the extent to which it has been " worked ". Such variations probably correspond quite helpfully with those occurring in a natural series of rocks of varied competency.

ACKNOWLEDGMENTS

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DESCRIPTION OF PLATE XI

FIG. 1.—Elevation of quartzite layer, 6 inches thick, emplaced within epidiorite (see Text-fig. 6, White Cow crush-belt) as a consequence of recumbent-fold thrusting towards the south-west. The quartzite layer underlies the part of the epidiorite, on the right of the photograph, which is part of the thrust mass. The thrust replaces and lengthens the inverted limb of an overfold which, as shown in the photograph, curls the basal quartzite layer over the part of the epidiorite which is seen on the left of the photograph. See Text-fig. 7.

FIG. 2.—" Fishes " of quartzite in mica-schist resulting from (*a*) small-scale folding and repeated shredding of the rocks by thrusting during the Older fold-period, and (*b*) subsequent mangling-out of the tectonic forms into the sheet-dip of the north-western limb of a Caledonoid anticline. The form in the centre-front of the photograph is the closure of a small-scale recumbent syncline, now isolated from its limbs. The section photographed, about 5 feet long, is exposed in an embayment to the south-west of the White Cow Rock, and is part of a steep face of the rocks underlying the epidiorite.

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FIG. 1.

FIG. 2.

TECTONIC STRUCTURES, WHITE COW ROCKS, MALIN HEAD, CO. DONEGAL.

Further Data on the Petrology of the Pelitic Hornfelses of the Carn Chuinneag–Inchbae Region, Ross-shire, with Special Reference to the Status of Almandine

By R. I. Harker

Abstract

The pelitic hornfelses surrounding the Carn Chuinneag–Inchbae complex have, with the other rocks in the area, been subjected to the regional metamorphism which affected the Moine Series as a whole. The minerals of the original hornfelses have reacted to form assemblages which are typical of regional rather than purely thermal metamorphism although the massive structure of the hornfelses is well preserved.

By studying the changes which took place in the hornfelses during the regional metamorphism it has been possible to deduce the most likely mineral assemblages indigenous to the original hornfelses.

The results of the chemical analyses of some of the pelitic hornfelses are presented and chemical data on some of the garnets are given. It is considered unlikely that all the garnets in the hornfelses, which are essentially almandines, formed as a result of the earlier contact metamorphism: some at least probably occur as a result of the later regional metamorphism.

Introduction

THE most comprehensive account of the Carn Chuinneag-Inchbae complex and its environment is to be found in the memoir of the Geological Survey of Scotland (1912), explaining the "one inch" sheet no. 93. In this memoir a considerable amount of information has been presented on the various rock types within the region. Orthogneisses outcrop as two large masses which are predominately granitic and are surrounded by rocks of the Moine series into which they were intruded prior to the regional metamorphism of the North-west Highlands. As a consequence of this earlier intrusive activity the crystalline schists of the area contain relics of undeformed hornfelses. The area thus includes a series of sediments which were partly hornfelsed by contact metamorphism after which both the hornfelses and the unmetamorphosed sediments of the series were regionally metamorphosed. In the field every stage between hornfelses and schists can be traced within a variable but usually comparatively short distance. All stages of transition can also be found between pelitic and psammitic rocks, through semipelitic horizons to typical Moine "granulites". In certain parts, and interbedded with the pelitic rocks, there are calcs-silicate bands.

The purpose of this paper is to supplement the data already published on the pelitic members of the series as it is from these that the metamorphic history of the region can most readily be interpreted.

Mineralogy

The bulk of the pelitic rocks, whether hornfelses or schists, are formed of the following minerals. Quartz, albitic plagioclase and biotite are ubiquitous, muscovite and garnet extremely common, whilst potash feldspar appears less frequently. Chlorite occurs rather sporadically and often as an alteration product of garnet. Iron ores, sphene, apatite and zircon comprise the more common accessories. The widespread occurrence of almandine in the pelitic schists indicates that the area as a whole was in the " almandine zone " of regional metamorphism.

Although in the pelitic hornfelses no andalusite nor cordierite has been found, pseudomorphs, often largely of kyanite, after both these minerals provide evidence of their previous occurrence (Tilley, 1935). The " chiastolites " in the hornfelses of Sgor a ' Chaoruinn are especially striking.

Table 1.—Analyses of Pelitic Rocks from the Carn Chuinneag Aureole

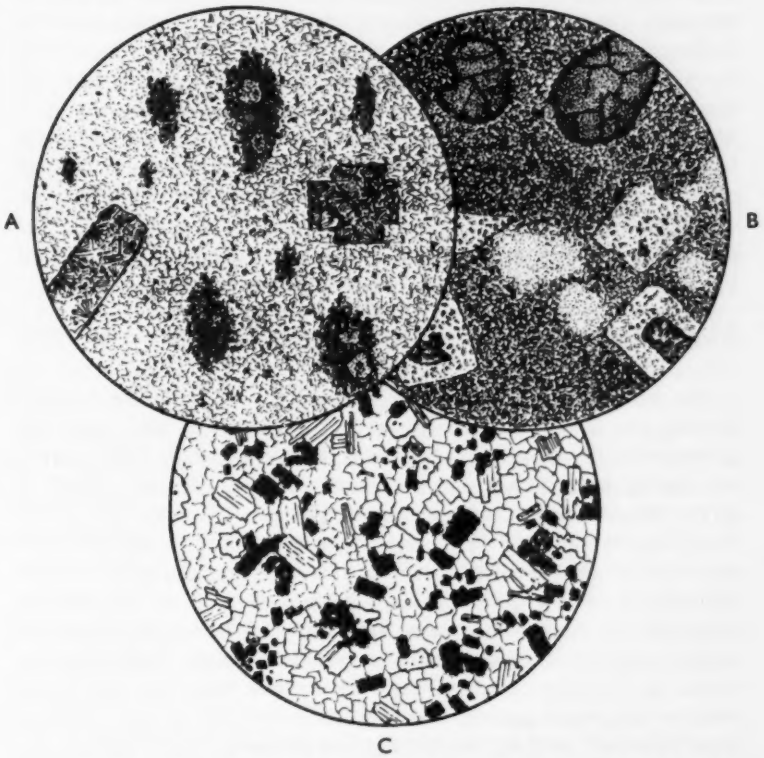
	1	2	3	4	5
SiO_2 . .	62·85	61·87	60·91	65·98	63·74
Al_2O_3 .	20·58	20·94	20·45	14·59	17·23
TiO_2 .	0·88	0·86	1·20	1·69	1·11
Fe_2O_3 .	1·10	1·06	1·40	0·31	0·94
FeO . .	5·10	4·76	5·29	5·19	5·03
MgO .	1·45	1·50	1·35	3·38	1·70
MnO .	0·04	0·05	0·03	0·09	0·24
CaO .	0·72	0·77	0·91	0·58	1·60
Na_2O. .	1·67	1·44	1·26	3·87	2·97
K_2O . .	4·06	4·32	4·06	2·88	3·55
P_2O_5 .	0·16	0·18	0·55	0·18	0·24
H_2O + .	0·97	2·24	2·21	0·90	0·98
H_2O − .	0·17	0·04	0·20	0·18	0·15
(Co, Ni)O .	n.d.	n.d.	n.d.	n.d.	0·04
FeS_2 .	n.d.	n.d.	n.d.	n.d.	0·32
	99·75	100·03	99·82	99·82	99·84

1. " Chiastolite " hornfels, see Text-fig. 1*a*, taken from 50 yards S. of summit of Sgor a'Chaoruinn.
2. Muscovite-rich garnetiferous hornfels with biotite and some recrystallized kyanite (parent rock of garnet, anal. 2, Table 4), ½ mile E. of Carn Sonraichte and in the Allt Sron Fhearchair.
3. Deformed " chiastolite " hornfels with muscovite, biotite, and garnet. Staurolite occurs in association with the kyanite of the pseudomorphs after andalusite. ½ mile N.N.W. of summit of Sgor a'Chaoruinn, i.e. at the W. end of Loch a'Chaoruinn.
4. " Cordierite " biotite plagioclase hornfels, see Text-fig. 1*c*. In the rock analysed, almandine is present in small amount (parent rock of biotite whose composition is given in Table 2), ½ mile S. of Mullach Creag Riaraidh and in the Diebidale River.
5. Garnetiferous biotite-hornfels, Garbh Allt, 1 mile S.E. of Glencalvie lodge (anal.: E. G. Radley, Geol. Survey. Memoir, 1912).

All the above rocks contain quartz and plagioclase in varying amounts.

The Chemically Analysed Pelitic Rocks

The chemical compositions of five hornfelsed pelitic rocks are given in table 1. The first three compare very closely with each other although they appear appreciably different in hand specimen and in thin section. The first is fine grained, slightly fissile and contains well developed " knoten " and pseudomorphs after chiastolite : its appearance in thin section is indicated by Text-fig. 1*a*. The second is extremely



Text-fig. 1 (*a*) " Chiastolite " hornfels (× 20), from 50 yards S. of summit of Sgor a'Chaoruinn. Brush-like aggregates of kyanite have replaced the andalusite. The dark spots are of biotite, kyanite needles, and garnet. The matrix is essentially of quartz, albite, and muscovite.

(*b*) " Chiastolite " hornfels (× 20), from Sgor a'Chaoruinn. Zoned garnets, pale " knoten " rich in muscovite and pseudomorphs of kyanite and shimmer aggregate after andalusite are set in a micaceous matrix containing quartz and albite-oligoclase.

(*c*) " Cordierite " hornfels (× 25), from ½ mile S. of Mullach Creag Riaraidh and in the Diebidale River. Biotite, kyanite needles, and shimmer aggregate have replaced cordierite. The other minerals illustrated are quartz, biotite, and albite. Small anhedral garnets also occur in this rock.

massive and is richer in muscovite and of coarser grain than the first ; it contains large spongy garnets whilst kyanite occurs in some quartz-rich regions of this rock and forms crystals about ½ mm. long. It is thought probable that this kyanite recrystallized from either original cordierite or andalusite and has become segregated to a lesser degree but by a similar tendency to that which caused the quartz-kyanite veinlets to develop (see Tilley, 1935, fig. 3, p. 95). No. 3 is phyllitic with a corrugated plane of foliation producing the lineation seen in the field. The pseudomorphs after andalusite are considerably distorted and staurolite accompanies the kyanite in them. The fourth contains less alumina than the other three and the kyanite in it is confined to the minute needles in the pseudomorphs after cordierite. Muscovite is absent and the large amount of albitic plagioclase is to be correlated with the higher ratio of Na_2O to K_2O than in the first three. A small amount of garnet occurs in this rock, which does not appear in Text-fig. 1*c*. The fifth analysis is of a garnetiferous biotite hornfels ; the analysis was carried out by E. G. Radley on behalf of the Survey and the rock has been described and figured in the Memoir (1912).

Reactions which Occurred in the Hornfelses during the Regional Metamorphism

The replacement, on regional metamorphism, of andalusite by both kyanite and shimmer aggregate, and of cordierite by the same two accompanied by biotite, has been described in detail by Tilley (1935) who has discussed the transient character of the kyanite which is absent from the surrounding schists proper. There is little doubt that all the kyanite now seen formed directly or indirectly from thermal metamorphic minerals, i.e. andalusite and cordierite, during the period of regional metamorphism (Tilley, 1935, p. 97). It is not difficult to imagine the andalusite and cordierite approaching a limit of stability as the conditions of regional metamorphism intensified. Such minerals might, in the presence of sufficient water, have been able to form a shimmer aggregate directly, and in many places this is their ultimate fate. However, until sufficient water was available, such a change was impossible. Under the relatively dry conditions which must have prevailed in the hornfelses, on completion of their thermal metamorphism, in contrast to the surrounding then unmetamorphosed sediments, kyanite was the more stable phase. As the hornfelses were progressively deformed, their impermeability was broken down and much of the kyanite was hydroxylated. Without denying that other factors may have played a part in facilitating the formation of kyanite in the hornfelses, but not in the schists, on regional metamorphism, it appears that the most significant factor was the difference in bulk composition with respect to the water content of the hornfelses (in

which kyanite formed) and of the sediments (which ended up as almandine schists). That it is possible for minerals characteristic of an anomalously high metamorphic " grade " to crystallize due to a restriction in the amount of water present has been suggested by Yoder (1952) from data obtained in the laboratory. There is probably a general similarity between the status of kyanite in the Carn Chuinneag-Inchbae aureole and that of hornblende in the low grade epidiorites described by Wiseman (1934), and in which, possibly owing to a deficiency of water at the onset of regional metamorphism, hornblende rather than chlorite is formed from the high temperature primary phases of the basic intrusives (Wiseman, 1934, p. 374).

In some sections, staurolite has been observed in association with the kyanite of the pseudomorphs after andalusite. Bosworth (1910, p. 396), also mentioned the occurrence of staurolite in the Carn Chuinneag region. Its presence may be the result of a somewhat unusual and localised metamorphic facies which prevailed at some stage in the history of the hornfelsed rocks for it has only been seen by the present writer in specimens taken from the region immediately to the west of Loch a ' Chaoruinn, three miles S.S.W. of Carn Chuinneag itself. A specimen from this locality, containing an appreciable amount of staurolite, was analysed but it shows no marked chemical peculiarities when compared with other staurolite-free pelitic rocks from the Carn Chuinneag region, see table 1. Other changes in the mineralogy of the original hornfelses, instigated by the later regional metamorphism, though less spectacular, undoubtedly occurred as evidenced, for example, by the difference in colour of the biotites (Hall, 1941). In the more massive hornfelses the biotite is often a red-brown colour, whilst in the schistose rocks it is yellow-brown. There is little evidence for any change in composition of the plagioclase but orthoclase has been sericitized. Quartz and muscovite, apart from that which actually took part in specific reactions involving other minerals, recrystallized.

The Original Mineral Assemblages in the Hornfelses

An attempt has been made to deduce the mineral assemblages of the original hornfelses before their dynamo-thermal metamorphism, though any inference on the exact nature of the phases which were present must be speculative. Sometimes it is easy to see that andalusite was present because of the preservation of excellent pseudomorphs. Less frequently is it possible to see that cordierite was present from such evidence. The original FeO : MgO ratios in the biotite and cordierite cannot be known and it is not possible to be sure what phase or phases are now represented by the different types of spheroidal " knoten ". By making certain tentative assumptions, however, it is possible to

draw up a scheme of mineral assemblages which probably existed in the hornfelses prior to their becoming involved in the regional metamorphism.

The more obvious pseudomorphs after cordierite consists of kyanite needles set in a matrix of fine grained biotite with a variable amount of shimmer aggregate, depending on the degree of alteration of the kyanite or of the original cordierite ; it has therefore been assumed that the shapeless " knoten " formed essentially of these same minerals represented original anhedral cordierite. Likewise, patches consisting only of kyanite in brush-like aggregates with or without shimmer aggregate, probably represent original andalusite.

In the past, it has been assumed that the garnet of the pelitic hornfelses developed during the formation of the thermal aureole (see Harker, 1932, p, 339), and Flett (Geol. Survey 1912) has stated that it appears that on regional metamorphism the garnet has, in certain instances, initially broken down to biotite. For the purpose of this reconstruction therefore the garnet will be considered as " thermal " (as opposed to " regional "), in order to examine its occurrence as such in the light of the mineralogical phase rule. As however in the view of the present writer there is some doubt as to the time of origin of at least some of the almandine in the hornfelses its status will be discussed more fully in a later section.

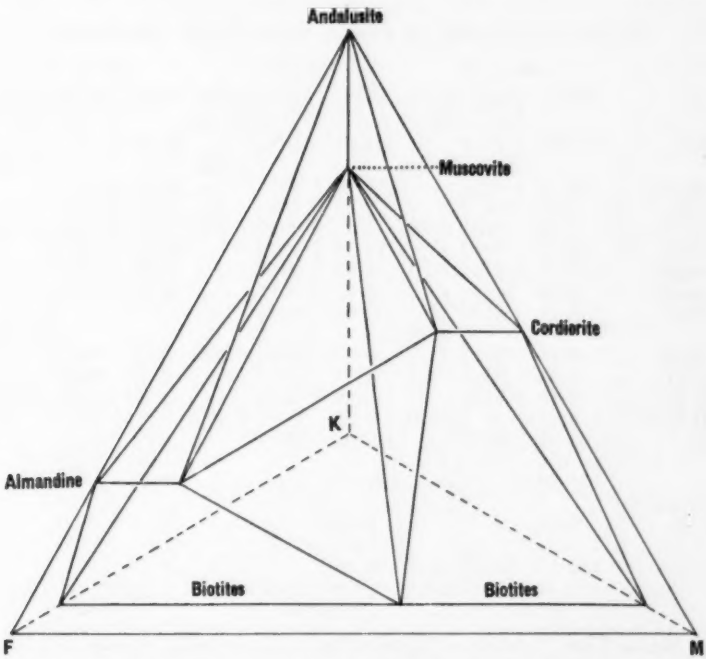
It appears probable that the majority of the original hornfelses which contained any four of the phases which are shown in the following A K F M diagrams, fall into four main groups as follows :—

(1) Andalusite + almandine + cordierite + muscovite + quartz ± plagioclase
(2) Biotite + almandine + cordierite + muscovite + quartz ± plagioclase
(3) Almandine + biotite + andalusite + muscovite + quartz ± plagioclase
(4) Cordierite + biotite + andalusite + muscovite + quartz ± plagioclase

Examples were also seen to contain
Potash feldspar + muscovite + biotite + quartz ± plagioclase and it is possible that,
Andalusite + almandine + cordierite + biotite + quartz ± plagioclase assemblages also occurred.

Quartz was present in all the sections examined and no silica-deficient minerals were seen. There is no evidence that regions of water deficiency were realized during the progressive thermal metamorphism of the pelites though the hornfelses were probably

more or less dried out eventually at the decline of the igneous activity. It is not unlikely therefore that water was present as a separate phase in addition to those now seen. As only the maximum number of possible phases which should normally have coexisted is being considered here, water must be regarded as having been present. The mineral assemblages of the rocks including the almandine + cordierite + muscovite hornfelses (i.e. 1 and 2 above) are represented schematically by the composition tetrahedra in an A K F M tetrahedron (see Text-fig. 2). This tetrahedron is developed directly from Eskola's (1915)



TEXT-FIG. 2.—The mineral assemblages of the almandine-cordierite-muscovite " facies ".

A K F diagram by the splitting up of the F component (FeO + MgO) into F and M (for FeO and MgO respectively) and it has been described and utilized by Barth (1936). It is necessary here to treat FeO and MgO separately because, in the phases under consideration, they behave to a large extent as separate components. Minerals in which solid solution occurs involving the isomorphous interchange of Fe and Mg (i.e. almandines, cordierites and biotites) are represented by lines : minerals whose compositions are fixed within narrow limits with respect to the components plotted are represented by points.

In accordance with the mineral assemblages inferred from observations of thin sections the tetrahedron is first divided into two parts by the muscovite-biotite plane. That part of it which now lies on the opposite side of this plane from potash feldspar is itself divided by the almandine-cordierite-muscovite plane.

The composition of the cordierite in equilibrium with almandine is unknown as no cordierite has been found. It is possible that the FeO : MgO molecular ratio may in some cases have been as high 1 : 1. However, iron-rich cordierites (as opposed to iron cordierites) when

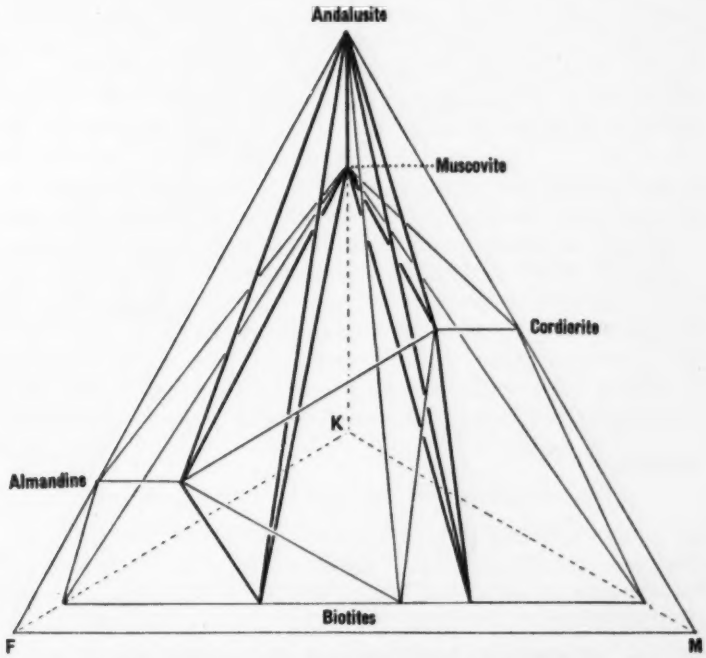
TABLE 2.—ANALYSES OF BIOTITE FROM PELITIC HORNFELSES

	Biotite 1			*Biotite* 2			*Biotite* 3		
	Wt %	Metal atoms: 12 (O, OH, F)		Wt %	Metal atoms: 12 (O, OH, F)		Wt %	Metal atoms: 12 (O, OH, F)	
SiO_2	36·47	2·711	4·000	34·87	2·665	4·000	35·21	2·669	4·000
Al_2O_3	19·70	1·722		19·79	1·782		20·07	1·792	
TiO_2	2·59	·143		5·12	0·294		2·65	0·151	
Fe_2O_3	1·44	·080	2·886	1·72	0·099	2·823	0·68	0·036	2·826
FeO	15·37	·954		17·79	1·137		16·88	1·064	
MgO	11·44	1·276		7·42	0·845		9·78	1·114	
MnO	0·07	·005		0·02	·001		0·35	·023	
CaO	0·17	·014	0·903	0·54	·044	0·940	tr	—	0·941
Na_2O	0·16	·028		0·67	·099		0·37	·054	
K_2O	8·98	·856		8·18	·797		8·94	·864	
H_2O+	3·58	1·782	1·800	2·89	1·472	1·573	3·80	1·918	1·918
H_2O-	0·15	—		0·29	—		0·53	—	—
F_2	0·08	·018		0·42	·101		n.d.		
	100·20			99·72			99·26		
Less O for F_2	0·03			0·18			—		
	100·17	S.G. = 3·01		99·54			99·26		
	γ = 1·642, reddish-brown.			1·661, reddish-brown.			1·638, orange rufous.		

1. Biotite from " cordierite "-biotite-plagioclase hornfels. Parent rock also analysed, see anal. 4, Table 1.
2. Biotite from garnet-spinel-cordierite-plagioclase-biotite-hornfels, Belhelvie, Aberdeenshire (Stewart, 1942).
3. Biotite from cordierite-biotite hornfels, Tenryukyo, Japan (Tsuboi, 1938).

found in metamorphosed argillaceous rocks are perhaps more usually associated with the higher grades of metamorphism, for example close to the contacts of basic igneous rocks (Tilley, 1924, p. 52, Read, 1929, and Stewart, 1942) or in sillimanite gneisses (Folinsbee, 1941). The point representing the composition of cordierite in equilibrium with almandine has, in view of the lack of any chemical evidence, simply been chosen conveniently (FeO : MgO approx. 1 : 3, mol. ratio) to clarify the A K F M diagram.

The point representing the composition of the biotite in equilibrium with cordierite and almandine together has been plotted from analysis of the biotite in a hornfels which contained a small amount of garnet and some pseudomorphs after cordierite, and which was collected from the north side of the Carn Chuinneag complex (see Table 2). Apart from having a slightly lower FeO : MgO ratio, this biotite is very similar in composition to that from a cordierite-biotite hornfels from Tenryukyo (Tsuboi, 1938). The biotite of the garnet-spinel-cordierite-



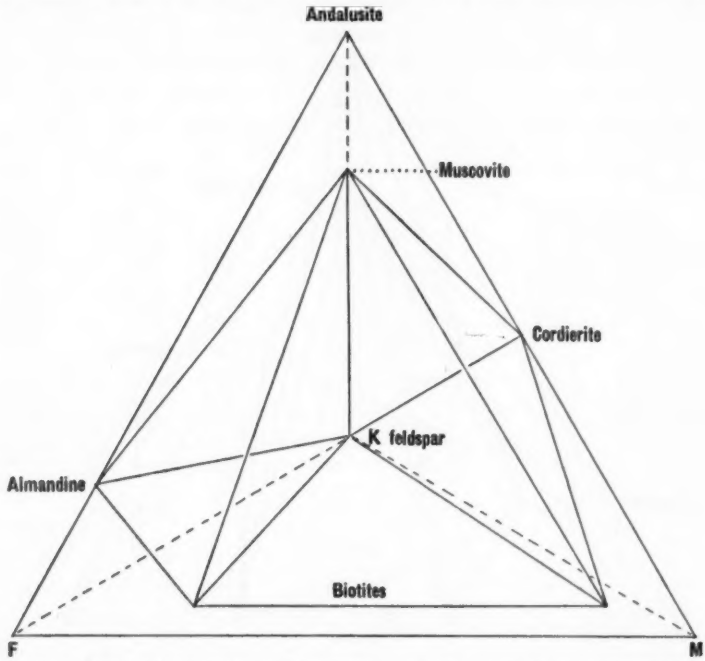
TEXT-FIG. 3.—The mineral assemblages of the biotite-andalusite " facies ".

biotite hornfels from Belhelvie (Stewart, 1942) is also comparable though it contains a somewhat greater amount of titania. All three biotites are remarkably alike both in their content of total alumina and in the amount of alumina necessarily substituting for silica if the Z group—$(OH,F)_4X_2Y_6$ (Z_8O_{20})—is to be made complete.

The garnet from an analysed pelitic hornfels (anal. 2, Table 1) has itself been analysed (anal. 2, Table 4). However, it is not at all certain that this garnet was in equilibrium with cordierite so the point limiting the solid solution of pyrope in almandine is probably at least slightly to the MgO side of that corresponding to the analysed garnet.

The essential feature about the rocks represented in Text-fig. 2, is that andalusite and biotite apparently did not co-exist.

In Text-fig. 3 the mineral assemblages in the hornfelses in which andalusite and biotite can occur together are represented. The thicker lines emphasize the almandine-biotite-andalusite-muscovite and the cordierite-biotite-andalusite-muscovite tetrahedra. The FeO : MgO ratios in the biotites of these two assemblages are represented by possible positions along the biotite solid solution line, i.e. on either side of the



TEXT-FIG. 4.—Almandine-potash feldspar and cordierite-potash feldspar assemblages.

point representing the biotite of the andalusite-almandine-cordierite-biotite assemblage.

Many of the hornfelses from the Carn Chuinneag-Inchbae area fall into two and three phase fields with respect to these diagrams, for example into the regions defined by muscovite-biotite, cordierite-biotite and biotite-muscovite-cordierite. As these, and several other assemblages can be recognized in both Text-figs. 2 and 3 they are less instructive than those which are characteristic of one or the other of the A K F M diagrams.

In both Text-figs. 2 and 3, it has been assumed that all the biotites

between and including annite and phlogopite would have been stable and would, if equilibrium conditions were realized, have formed in rocks of the appropriate composition. In the event of annite and phlogopite being unstable, then it is possible that two new composition tetrahedra would appear in the diagrams. These would be bounded by almandine-biotite-muscovite-potash feldspar and by cordierite-biotite-muscovite-potash feldspar respectively (see Text-fig. 4). Such assemblages as these were not seen in the Carn Chuinneag-Inchbae rocks but might occur in rocks sufficiently rich in potash and containing, in the first case more iron, and in the second case more magnesia, than could be absorbed into the biotite structure under the conditions of crystallization. Both these fields would, of course, vary in size and shape with every small change of P T conditions and of the concentrations of the minor elements, as indeed would all the fields involving solid solution.

It was observed that the distribution of the deduced mineral assemblages in the original Carn Chuinneag-Inchbae aureole was not entirely at random. Assemblages nos. (1) and (2), represented by Text-fig. 2, were characteristic of the hornfelses adjacent to the north-eastern part of the complex whilst (3) and (4) represented by Text-fig. 3, were characteristic of the hornfelses to the south-west. The significance of this is not yet clear, chiefly owing to the rather uncertain status of the garnet, but it would suggest that somewhat different P T conditions may have prevailed in the two regions during the thermal metamorphism.

The Occurrence of Garnet in the Hornfelses

As mentioned earlier, the garnet in the pelitic hornfelses has, in the past, been regarded as having formed during the thermal metamorphism. It is clear, however, that several phase changes were induced in the mineralogy of the hornfelses by the regional metamorphism before any appreciable strain was imposed on them. Such changes are evident from the partial or complete disappearance or appearance of some phases or in the case of biotite, from a change in colour. It is, therefore, not out of place to re-examine the status of the garnet and to consider whether in fact that which is now so common in the hornfelses, formed during the earlier period of thermal metamorphism or during the later period of regional metamorphism.

The garnet occurs with approximately equal abundance in both the schists and the hornfelses. It is in the former that the largest crystals are to be found, but crystals up to 0·75 cms. in diameter are not uncommon in the latter. In both the schists and the hornfelses the garnets are usually sub- to an-hedral and riddled with inclusions : they frequently show signs of alteration to chlorite and sometimes to biotite.

In the schists, the inclusions in the garnets are occasionally arranged in spirals, indicating a paratectonic crystallization. Such rolled garnets are peculiar to the schists and absent from the hornfelses.

In many of the spotted hornfelses, the garnet is confined to the "knoten" which are largely composed of biotite and kyanite with a variable amount of shimmer aggregate (Text-fig. 1*a*). Furthermore, in a few examples, it has been possible to identify kyanite needles as inclusions in these garnets when the needles pervade the garnets in the same way that they do the rest of the "knoten". In a few of the hornfelses the garnets can be seen to be faintly but sharply zoned and to have distinguishable rims and cores (Text-fig. 1*b*). This zoning is apparent by a sudden change in the density of distribution of minute black opaque inclusions; sometimes, a slight colour change also occurs when the core is very faintly pink whilst the rim is grey. Within the limits of accuracy of measurement (± ·003) no difference in refactive indices could be detected between the cores and the rims.

The samples of garnet from the Carn Chuinneag-Inchbae hornfelses on which chemical data are available can be seen to contain a high percentage of almandine and only a low percentage of spessartine, in spite of the concentration in the garnets of the manganese available in the rocks (Table 3). It may be noted that analysis no. 2 in this table

TABLE 3.—PARTIAL ANALYSES OF GARNETS FROM THE HORNFELSES OF THE CARN CHUINNEAG–INCHBAE REGION

	1	2	3
Wt % FeO .	32·40	37·31	34·85
Wt % MnO .	0·72	0·46	0·67
FeO/MnO .	45	81	52
Wt % Almandine .	74·7	86·0	80·9
Wt % Spessartine .	1·7	1·1	1·5
MnO in parent rock	0·08*	0·05	0·05
$\frac{\text{MnO in garnet}}{\text{MnO in rock}}$.	9·0	9·2	13·4
R.I. of garnet .	n.d.	1·809	1·805

1. Garnet from muscovite-biotite-garnet hornfels. S.E. side of Carn Chuinneag aureole. (Tilley, 1935.)
2. Garnet from "chiastolite" hornfels with garnet. Sgor a Chaoriuinn.
3. Garnet from muscovite-garnet-hornfels with subsidiary biotite. ½ mile E. of Carn Sonraichte and in the Allt Sron Fhearchair.

* Anal.: R. I. Harker.

is of one of the zoned garnets referred to above and illustrated in Text-fig. 1*b*. The outer rims of these garnets are relatively narrow so that the greater part of the material analysed represents the cores. It was impossible to separate out the minute opaque inclusions in the time available but from a rough, though liberal estimation of their amount it was concluded that they could not account for more than 3% of the FeO in the analysis, assuming that they were entirely of magnetite.

In fact some of them may well be of ilmenite or carbon (a few were dissolved out in hydrochloric acid and found to contain total iron, as $Fe_2O_3 : TiO_2 : MnO$ in the ratio of 127 : 19 : 1.).

TABLE 4.—ANALYSES OF HORNFELSES AND ALMANDINES

	Carn Chuinneag		Gadernheim		Belhelvie		Bohemian Massif		Botallack	
	1	2	3	4	5	6	7	8	9	10
SiO_2	61·87	36·90	47·76	38·63	40·77	36·56	43·46	36·60	53·55	35·58
Al_2O_3	20·94	21·93	24·59	21·81	25·90	21·07	17·38	20·47	8·40	21·94
Fe_2O_3	1·06	0·40	3·14	1·47	1·02	0·47	5·65	1·80	0·74	none
TiO_2	0·86	0·05	1·61	—	2·13	0·51	0·56	0·54	0·40	tr.
FeO	4·76	34·85	10·10	29·38	17·52	33·51	19·22	33·50	26·75	38·54
MnO	0·05	0·67	—	1·90	0·24	1·81	0·28	0·92	0·25	0·70
MgO	1·50	2·99	2·79	4·81	4·95	3·97	2·76	4·18	2·37	0·68
CaO	0·77	2·58	3·68	1·57	2·07	1·61	1·92	1·72	1·20	1·68
Na_2O	1·44	—	2·23	—	1·09	—	2·88	—	0·26	—
K_2O	4·32	—	1·68	—	2·73	—	2·38	—	2·94	—
P_2O_5	0·18	—	0·41	—	0·02	—	0·58	—	0·06	—
SO_3	—	—	0·48	—	—	—	—	—	—	—
$H_2O +$	2·24	—	1·38	—	0·97	—	1·92	0·66	2·15	—
$H_2O -$	0·04	—	0·12	—	0·26	—	1·03	—	0·65	0·12
	100·03	100·37	99·97	99·57	99·67	99·51	100·02	100·39	99·72	99·24
Weight % of										
Almandine		80·87		72·8		77·85		78·3		91·33
Spessartine		1·50		4·8		4·24		2·2		1·69
Pyrope		10·13		17·3		13·26		14·7		2·35
Grossularite		5·45		0·2		3·04		3·8		4·63
Andradite		2·05		4·9		1·61		1·0		—

(1) Muscovite–rich garnetiferous hornfels with biotite and a small amount of recrystallized kyanite. Parent rock of garnet anal. 2 below. ½ mile E. of Carn Sonraichte and in the Allt Sron Fhearchair, Carn Chuinneag–Inchbae aureole.
(2) Almandine from hornfels from Carn Chuinneag–Inchbae aureole.
(3) Garnetiferous hornfels. Parent rock of (4), from Gadernheim, Odenwald. G. Klemm (1919).
(4) Almandine from hornfels, from Gadernheim, Odenwald. G. Klemm (1919).
(5) Garnet-spinel-cordierite-plagioclase-biotite-hornfels. Parent rock of (6). F. H. Stewart (1942).
(6) Almandine from Belhelvie, Aberdeenshire. F. H. Stewart (1942).
(7) Garnet-biotite-cordierite-hornfels. Parent rock of (8). K. Preclik (1924).
(8) Almandine from the Bohemian Massif. Locality unknown. K. Preclik (1924).
(9) Grunerite-biotite hornfels. Country rock to which (10) belongs. C. E. Tilley (1935*a*).
(10) Almandine from Botallack. A. R. Alderman (1935).

So far, there is no evidence from their chemical composition that the garnets originated as a result of contact metamorphism for they contain appreciably less manganese than most garnets produced by contact metamorphism (Tilley, 1926, p. 47). There is evidence, however, that a high manganese content is not always necessary for the formation of garnet during thermal metamorphism and almandines have been found in unstrained hornfelses elsewhere. The chemical compositions of

some of these and their associated rocks are given in Table 4 which includes that of the most fully analysed garnets from the Carn Chuinneag region for comparison. All are essentially almandines, poor in manganese, calcium and ferric oxides. The amount of MgO in all these garnets, except in the specimen from Botallack, is of the same order of magnitude. However, the proportion of pyrope is characteristic of that found in many garnets from biotite schists and is appreciably less than that frequently found in rocks which are believed to have formed at greater depths in the earth's crust, e.g. in eclogites (Wright, 1938).

It has been conjectured that if contact metamorphism is effected under a sufficiently high hydrostatic pressure then almandine may form in pelitic hornfelses (Harker, 1932, p. 55). Many attempts by various workers to synthesize almandine have failed, and this failure was most probably due to the experiments being carried out under insufficient pressure. Schairer and Yagi (1952) for example found that at the lowest temperature at which they could, in reasonable time, crystallize a glass of almandine composition, iron cordierite, fayalite and hercynite crystallized instead of almandine. It is probable that a sufficiently high pressure would have greatly favoured the formation of almandine from the glass because of the volume change shown by the following equation :—

$2Fe_2Al_4Si_5O_{18}$	+	$5Fe_2SiO_4$	+	$FeAl_2O_4$	=	$5Fe_3Al_2Si_3O_{12}$
(cordierite)		(fayalite)		(hercynite)		(almandine)
460 cc.		240 cc.		40 cc.		600 cc. approx.

Volume reduction of approximately 19 per cent.

In the presence of excess silica as quartz the volume reduction is only slightly less :—

$Fe_2Al_4Si_5O_{18}$	+	$2Fe_2SiO_4$	=	$2Fe_3Al_2Si_3O_{12}$	+	SiO_2
(cordierite)		(fayalite)		(almandine)		(quartz)
230 cc.		90 cc.		240 cc.		23 cc. approx.

Volume reduction of approximately 18 per cent.

Coes's synthesis of almandine, among many other minerals which previously defied synthesis, indicates beyond all reasonable doubt that high pressure greatly favours its formation (personal communication). It is probable, therefore, that the formation of a thermal aureole at considerable depths, might involve the development of almandine.

Consideration of Petrographic and Chemical Data : Conclusions

It seems reasonable to suppose that the zoned garnets which appear to be essentially almandines, formed in two distinct stages. The cores may represent a generation of garnets produced by the contact

metamorphism whilst the rims could have developed during the later regional metamorphism. It is noteworthy that so far zoned garnets have only been found in the hornfelsed rocks, i.e. not in the schists outside the aureole. If the cores formed during the thermal metamorphism then it is also possible that some of the apparently unzoned garnets are partly or wholly of this first generation.

Flett's interpretation that at the onset of regional metamorphism some of the garnet altered to biotite suggests that the garnets were in the hornfelses prior to the regional metamorphism. Such an alteration would probably occur in the event of the garnet in question having formed during the period of thermal metamorphism. However, it is also a probable retrograde effect, examples of which have been found in the schistose rocks of the area, although the more common product is chlorite.

The manner in which it is possible to draw up a scheme representing the original mineral assemblages in the garnetiferous hornfelses provides some circumstantial evidence suggesting that garnets may have developed during the contact metamorphism. In the majority of the garnetiferous hornfelses, the garnet appears in such a way as not to violate the phase rule.

On the other hand, sections have been seen in which there would appear to be one phase too many according to the mineralogical phase rule. These sections contained well developed pseudomorphs after andalusite set in a fine grained matrix containing biotite and muscovite : large anhedral " knoten " were present which originally may well have contained cordierite or chlorite but now contain garnet. It is, of course, not impossible that five phases in the A K F M diagram could have crystallized together in purely thermally altered rocks : neither does it unequivocally indicate disequilibrium. It is possible that if the hydrostatic pressure was high enough some of the garnetiferous " knoten " may have been initiated by the very early formation of almandine. This would leave the rest of the rock poor in FeO—especially in the immediate neighbourhood of the garnet round which cordierite might then be encouraged to develop.

Alternately, the garnets which occur in some of the " knoten " may, with the biotite and kyanite (or shimmer aggregate) accompanying it, be a direct alteration product from original cordierite. These garnets, however, often form up to at least an eighth of the volume of " knoten " of which the patches are a section. The following equation, showing the breakdown of cordierite to almandine, is one which might well be driven from the left towards the right at the onset of regional metamorphism. It involves a volume decrease of approximately 17% and much of the aluminum changes from four-fold to six-fold coordination.

$$\underset{\text{(cordierite)}}{12MgFeAl_4Si_5O_{18}} + \underset{\text{(muscovite)}}{7(OH)_2Al_3Si_3O_{10}K} = \underset{\text{(biotite)}}{7(OH)_2Al\,Si_3O_{10}K.Fe_3Mg_{12}}$$

2,800 cc. 1,000 cc. 1,070 cc.

$$+ \underset{\text{(kyanite)}}{30Al_2SiO_5} + \underset{\text{(almandine)}}{Fe_3Al_2Si_3O_{12}} + \underset{\text{(quartz)}}{27\,SiO_2}$$

1,350 cc. 120 cc. 610 cc. approx.

The approximate FeO : MgO ratio in the biotite has been taken from the analysis in Table 2. The analysed biotite and the biotite in the garnetiferous " knoten " are both of the same colour : the refractive index γ of a biotite from a typical garnetiferous " knoten " was found to be $1{\cdot}650 \pm {\cdot}002$ and so is slightly greater than that of the analysed biotite of which $\gamma = 1{\cdot}642 \pm {\cdot}002$. Muscovite has been taken as the most probable source of potash as it is a mineral quite likely to have been intimately associated with the original cordierite. The other obvious choice is potash feldspar but this has not been found with possible pseudomorphs after cordierite in the rocks of the region under discussion. It can be seen that even if the FeO : MgO molecular ratio in the cordierite is as high as 1 : 1, it would take approximately 22 ccs. of cordierite to produce 1 cc. of almandine. This ratio is too high to account for all the garnet seen in many " knoten " and it is considered more likely that the " knoten " provided suitable sites for the initial development of garnet at the onset of the regional metamorphism. At any rate, in those garnetiferous " knoten " in which both the garnets and the " knoten " as a whole contain kyanite needles as inclusions, the order of formation most probably was (1) cordierite (2) kyanite (3) garnet.

If garnet could develop in the " knoten " of the hornfelses during their regional metamorphism, it is probable that it would also have developed outside the " knoten " and in the hornfelses at the same time. There can surely be no doubt that the garnet in the surrounding schists developed during the regional metamorphism so that its simultaneous formation in the hornfelses is, from a consideration of the metamorphic facies, less remarkable than the development of kyanite from andalusite and cordierite. The objection might be raised that the components of the garnet would have difficulty in diffusing together to form this mineral in the impermeable hornfelses which may have had most or all of their interstitial water driven off at the close of their thermal metamorphism. Indeed a sufficiently active intergranular fluid might have completely suppressed the development of kyanite, although once this highly refractory mineral has formed, it powerfully resists decomposition and is itself capable of becoming segregated in the quartz-kyanite veinlets. Almost certainly, however, biotite and perhaps chloritic minerals contributed to the formation of garnet in the hornfelses : such reactions would involve the liberating of water locally where it would be most needed to facilitate diffusion and crystallization.

Whereas, it is felt that there is stronger evidence to support the suggestion that the garnets in certain hornfelses did not develop until the later period of regional metamorphism than there is to support their purely thermal development in other hornfelses, there is a large " no-man's land " from which any evidence as to their time of origin is absent. It is concluded that it would be unwise to correlate the formation of all garnets in the undeformed hornfelses with the thermal metamorphism.

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A Xenolithic Monchiquite Dyke Near Glenfinnan, Inverness-shire

By G. P. L. Walker and J. V. Ross

Abstract

The remarkable xenolithic dyke described in this paper is one of the little-known swarm of camptonite and monchiquite dykes of probable Permian age in the Northern Highlands of Scotland. It is the first intrusion containing xenoliths of ultrabasic rock to be recorded from the Glenfinnan area, although such xenoliths have been described from vents or intrusions of probable or known Permian age in Caithness in the extreme north of Scotland, and in Ayrshire and elsewhere in the south of Scotland.

Introduction

THE xenolithic dyke is seen in a deep gully high up in the corrie on the south side of the peaks Streap (2,988 ft.) and Streap Comlaidh, 4½ miles N.E. of Glenfinnan railway station in Inverness-shire, Scotland,[1] and 11 miles W.N.W. of Fort William. The dyke is poorly exposed at the head of the gully where it trends about N. 50° E. and cuts Moine granulites of high metamorphic grade. Associated with this dyke are others that are not xenolithic. One of these, near the lower end of the gully, is intruded into a shear-belt in the Moine erosion along which has caused the gully. This dyke is a few feet wide, trends N. 25° E. and is in places very amygdaloidal. Another dyke, or a continuation of the same one, occurs a little lower down ; it is three feet wide and trends N. 70° E. Near the head of the gully are other exposures of dyke-rock.

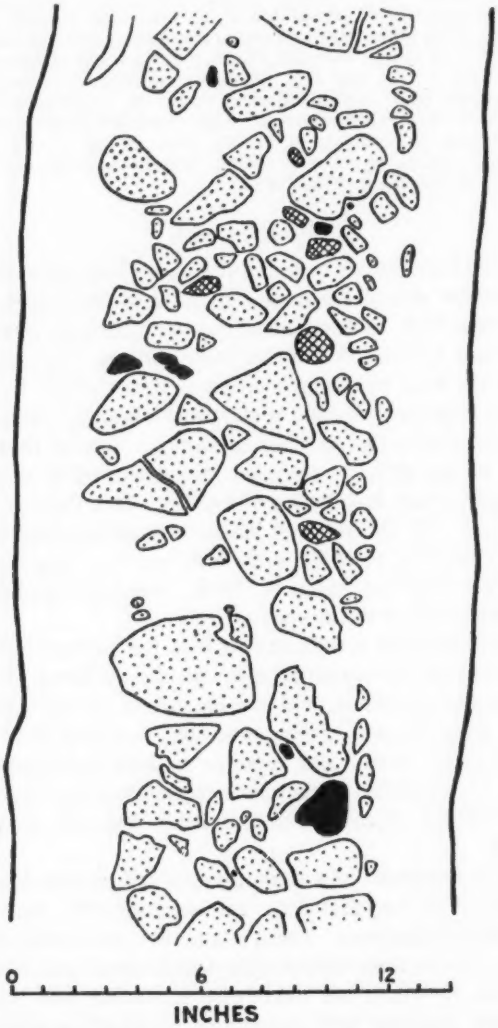
The head of the gully is difficult of access, but a considerable number of large blocks of the xenolithic dyke are strewn along the gully and in the river gravels below, and the characters of the dyke are best studied in these blocks. They show that it varies from twelve to twenty inches in width, and presents a sharp and chilled contact against the Moine rocks. Adjacent to the contact the Moine country-rock has suffered thermal metamorphism, and the dyke is firmly welded to it.

The dyke is composed of a dark, fine-grained basaltic-looking rock. Occasionally it is amygdaloidal, and exceptionally the amygdales reach one inch in diameter. There is an outer, non-xenolithic selvage to the dyke, two or three inches wide, which passes into a remarkable inner portion in which the dyke-rock is packed with xenoliths of various types, together with xenocrysts, some of considerable size.

[1] Latitude 56° 55′ N., longitude 5° 22′ W. National Grid Reference 17/951858.

The xenoliths make up more than fifty per cent of this inner portion (Text–fig. 1).

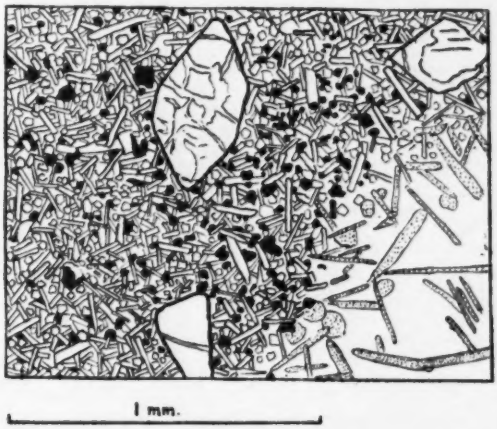
The great majority of the xenoliths are of a rather uniform peridotite. Associated are other types, including coarse-grained pyroxenites, banded feldspar-pyroxene rocks, and metamorphic rocks of Moine



TEXT-FIG. 1.—Sketch of part of the xenolithic dyke, from a block, showing xenoliths of peridotite (stippled), pyroxenite (cross-hatched), and schist (not ornamented), and xenocrysts (black).

type. The xenoliths are subangular and vary in size from a fraction of an inch to several inches; the largest seen measured eight by four inches. Some are cut by narrow veinlets of the dyke-rock. There is little sign of a preferred orientation of the xenoliths, apart from a slight tendency for elongation normal to the dyke walls.

Xenocrysts of pyroxene are quite abundant, rounded or irregular in shape, and often over an inch or even over two inches in diameter. Two types are seen, the one black and pitch-like which is titanaugite, and the other dull and green in appearance which is a diopsidic augite. Rare xenocrysts of feldspar are also seen, and very rare biotite.



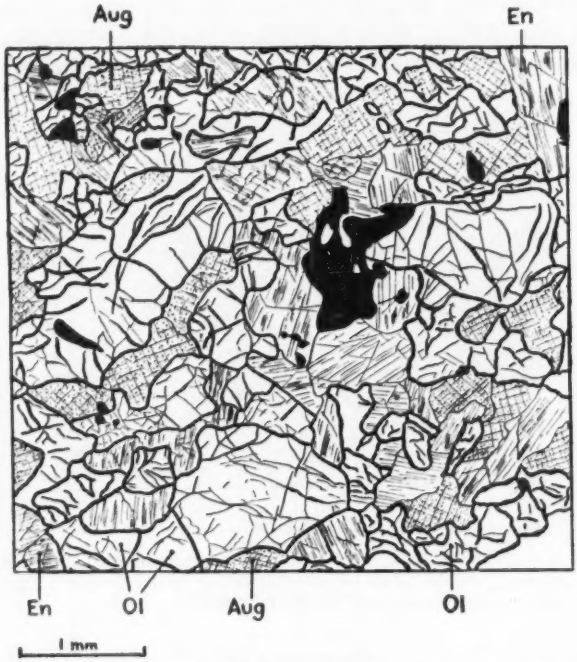
TEXT-FIG. 2.—Drawing (from a photograph) of monchiquite showing olivine phenocrysts and one of the numerous ocelli (bottom right) containing hornblende and biotite (stippled).

PETROGRAPHY

The dyke-rock.—The dyke-rock is a fine-grained monchiquite or olivine-fourchite (Text-fig. 2). Phenocrysts of olivine, normally ½ to 1 mm. in diameter, make up 20 per cent of the rock, together with sparse microphenocrysts of titanaugite. Titanaugite constitutes about three-quarters of the groundmass, the remainder including abundant opaque iron oxide and interstitial analcite, and a little biotite and hornblende.

The olivine phenocrysts are usually quite well-formed, but some rounding and corrosion is often present. There is a small amount of serpentinization. The crystals show no sign of zoning, and the optical properties ($\beta = 1{\cdot}677$) indicate about Fa 10. There are occasional minute inclusions of picotite. Titanaugite occurs mostly as idiomorphic prismatic crystals in the groundmass of the rock, which seldom

exceed 0·1 mm. in engtth. The prisms appear to show no preferred orientation, but often form small radiating clusters. Opaque iron oxide is abundant as small granules, with occasional small crystals of olivine and sparse little flakes of red biotite and prisms of brown amphibole. There is a colourless, isotropic base, probably analcite, interstitial to the crystals and often accompanied by chlorite and calcite.



TEXT-FIG. 3.—Section of lherzolite. Ol = olivine; En = enstatite; Aug = augite. Picotite shown black.

The monchiquite contains numerous coarser-grained patches or ocelli about one mm. in diameter. Growing into these are flakes of deep brown, strongly pleochroic biotite, prisms of dark brown barkevikite, and occasional prisms of titanaugite, larger than those in the body of the rock. There is much analcite in the ocelli, and sometimes it is accompanied by minute crystals of feldspar. Skeletal iron oxide is abundant. The ocelli pass into amygdales infilled with chlorite or calcite.

The dyke is chilled against the Moine granulites to a distance of several inches from the contact. The contact with the country-rock is sharp, and occasional minute veinlets of chilled monchiquite penetrate

the granulites. The Moine rocks, which are here quartz-feldspar-biotite granulites, have been thermally altered, and two main effects are observed. Firstly, there is the breakdown of biotite to a brown, amorphous mass, associated with iron oxide and a colourless mineral, possibly prehnite. Secondly, there is the development of numerous minute ramifying veinlets of felsitic material containing amygdale-like patches of chlorite, calcite, and chalcedony. These felsitic veinlets have been produced by fusion of the country-rock.

Lherzolite xenoliths.—The predominant type of xenolith is of a coarse-grained granular rock with allotriomorphic texture (Text-fig. 3) and having a modal composition [1] of : olivine 45, augite 23, enstatite 28, picotite 4. In the hand-specimen these minerals are readily distinguished, being respectively almost colourless, green, yellow, and black. The rock would perhaps best be called a lherzolite, although poorer in olivine than the type lherzolite.

The olivine occurs as colourless, more or less rounded crystals averaging about ½ mm. in diameter, and the optical characters ($2V\gamma = 84°$, $\beta = 1{\cdot}671$) indicate a composition about Fa 10. The augite is a pale green diopsidic variety with $2V\gamma = 60°$ and $\beta = 1{\cdot}690$. The crystals are somewhat larger than those of olivine, which they are often seen partially to enclose. Enstatite forms somewhat poikilitic crystals which tend to enclose the olivine and augite. It is very pale yellow-green and is not visibly pleochroic in sections of normal thickness. Its optical characters ($2V\gamma = 88°$, $\beta = 1{\cdot}675$) indicate an enstatite containing about 10 per cent of $FeSiO_3$. Picotite is an abundant accessory, as sub-poikilitic crystals moulded on the other minerals. In addition there is a very small amount of magnetite.

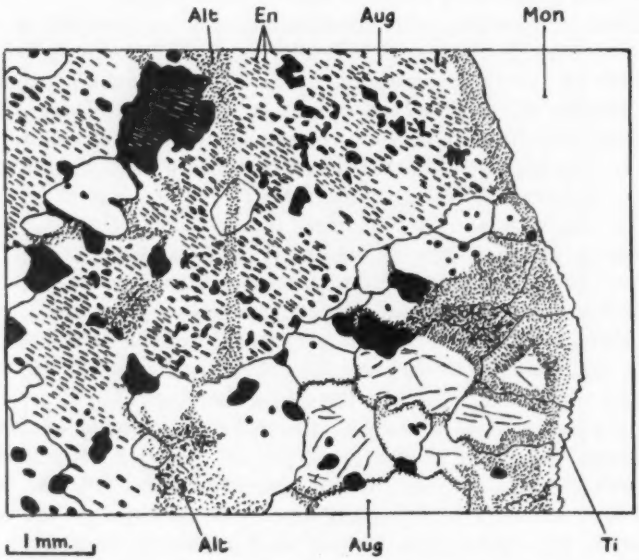
The contact with the monchiquite of the dyke is perfectly sharp, and the dyke appears to have had no effect on the minerals of the xenolith, all of which are perfectly fresh. The monchiquite itself is sometimes slightly chilled at the contact.

Pyroxenite xenoliths.—Some xenoliths are of rocks composed entirely of enstatite and augite. One (Text-fig. 4) consists of about 75 per cent augite and 25 per cent enstatite. Augite crystals up to one cm. in diameter contain oriented lamellae of enstatite and also round grains with the same optical orientation as the lamellae but, unlike them, extending beyond the boundaries of the augite. Other crystals of augite are smaller and lack the oriented lamellae, but often enclose round grains. Occasionally the enstatite contains regular film-like oriented lamellae of augite. Both minerals are almost colourless in thin section. To a depth of about ½ mm. from the contact with the

[1] Modes measured on thin sections by a point counter, and expressed in volume per cent.

monchiquite of the dyke the augite presents a fritted appearance due to partial fusion, and there is a later rim of fresh titanaugite deposited from the monchiquite in crystallographic continuity with the augite of the xenolith. In addition to the marginal fritting, alteration has proceeded inwards along the crystal boundaries.

Another xenolith consists of approximately 75 per cent enstatite and 25 per cent augite, and again both minerals are almost colourless.



TEXT-FIG. 4.—Drawing of part of a pyroxenite xenolith. Aug = augite; Alt = altered or partly fused rock, represented by close stippling; En = enstatite lamellae; enstatite is black, and most of that in the field has the same orientation; Mon = monchiquite; Ti = titanaugite rim at contact of xenolith and dyke.

Irregular but more or less equidimensional grains of enstatite averaging 1 to 2 mm. in diameter contain regular film-like lamellae and irregular blebs of augite, both with the same optical orientation. Apart from the lamellae, augite is mainly interstitial to the enstatite and sometimes has the same optic orientation as the inclusions in a neighbouring enstatite.

Xenoliths of Moine Type.—Under this heading come rather sparse xenoliths which probably represent metamorphic rocks of the Moine series, into which the dyke has been intruded. These rocks are medium- to coarse-grained granulites of Moine type and consist essentially of andesine and strained quartz, with rare sillimanite.

Numerous lanes of felsitic material which penetrate the xenoliths

along the grain boundaries are the product of partial fusion of the xenoliths. Adjacent to these lanes the plagioclase is corroded, and a marginal portion of the crystals is clouded by partial fusion to a colourless glass. The felsitic material contains abundant small crystals and microlites of alkali-feldspar, together with a good deal of chlorite and other secondary minerals. In places there are bundles of little needles of fibrolite or mullite, associated with minute octahedra of a grey or purple spinel, and one xenolith contains intergrowths of pyroxene with a deep green spinel. Near the margin of the xenoliths there has been considerable introduction of material into the felsitic lanes from the monchiquite of the dyke, and pale green augite or titanaugite, barkevikite, biotite, and skeletal iron oxide are developed. In one thin section the granulite contains zircon crystals enclosed by quartz and feldspar near the grain boundaries; similar grains of zircon are now found unchanged in the felsitic veinlets. Xenocrysts of quartz which have floated off into the monchiquite are sometimes mantled by a thin reaction rim of prisms of pale green augite, with some calcite.

Xenoliths of Pyroxene-gneiss.—Several thin sections of xenoliths of coarse-grained banded rocks have been examined which prove to contain pyroxene associated with plagioclase and, in some cases, perthite. Although some variation is shown by these rocks, they are clearly closely related to one another. Pyroxene constitutes from one quarter to nearly one half of the rocks and forms crystals averaging one to two mm. in diameter, grouped in spindle-shaped clusters a few mm. wide and oriented in one direction. The pyroxene is rather rich in iron and is either a ferrosalite or clinohypersthene. The plagioclase feldspar is a medium- to basic-andesine (in one rock it is oligoclase), as anhedral and more or less equidimensional grains that are little if at all zoned, and which show albite twinning.

One section, cut from a xenolith several inches long, has a green, pleochroic clinopyroxene with $2V\gamma = 58°$ and $\beta = 1{\cdot}708$, these characters indicating a salite or ferrosalite. The rest of the rock consists of roughly equal amounts of perthite and medium andesine that is slightly zoned. There is also a small amount of a very irregular antiperthite in which the potash feldspar is replacing the plagioclase. The perthite is a regular vein perthite in which the two feldspars are in almost equal amount. The feldspars form a granular aggregate with average grain-size of $\frac{1}{2}$ to 1 mm. A second xenolith has ferrosalite associated with magnetite, and plagioclase (andesine-labradorite) is the only feldspar. A third contains clusters of iron-rich, diallage-like augite with oriented lamellae, set in a granular aggregate of medium oligoclase, rarely antiperthitic, and there is accessory magnetite and apatite, the latter in large crystals interstitial to the feldspar. Particu-

larly interesting is one xenolith containing clinohypersthene. The mineral has the optical characters of a hypersthene containing about 30 per cent of $FeSiO_3$ ($\alpha = 1{\cdot}702$, $\beta = 1{\cdot}706$, $\gamma = 1{\cdot}718$, pleochroism X . . . pinkish-brown, Y . . . pale brown, Z . . . green) but in thin section most of the crystals give oblique extinction, usually between 10 and 25 degrees, and a study with the universal stage confirms that the axes of the indicatrix are inclined to the crystallographic axes.

Discussion

In western Inverness-shire there is a swarm of predominantly east–west dykes, mainly monchiquites and camptonites, and there is little doubt that the Streap xenolithic dyke belongs to this swarm. Other monchiquites in the Glenfinnan area resemble it very closely, except in their being non-xenolithic. The age of this swarm of dykes and of similar dykes in the Orkneys, Caithness and Sutherland, and further south in Colonsay has been fully discussed by Richey (1939, pp. 416–19), and more recently summarized by the Geological Survey (Phemister, 1948), the conclusion being reached that they are probably Permian. Direct evidence in Caithness indicates an age later than the Upper Old Red Sandstone, and in Morvern dykes of the swarm cut quartz-dolerite dykes probably of Permo-Carboniferous age. Negative evidence, in the apparent absence of known dykes of the swarm intersecting Mesozoic sediments, suggests a pre-Jurassic age, but it is still possible the dykes are Tertiary. However, age determination by the helium method indicates that dykes of similar petrographic character (but N.W. trend) in Colonsay are Permian. Further strong evidence is supplied by the close comparison of the monchiquites and camptonites with known Permian dykes and plugs in the Midland Valley of Scotland, one of which contains carbonated xenoliths of peridotite.

The genetic significance of the xenoliths in the Streap dyke may now be considered. Both the dyke-rock and the bulk of the xenoliths are thoroughly basic or ultrabasic in character, and both consist largely of olivine and pyroxene. There are obvious differences : for example, the titanaugite of the dyke-rock is represented in the xenoliths by a diopsidic augite and enstatite, and no sign of the alkaline characters of the dyke is found in the ultrabasic xenoliths, but the syngenetic relationship between these xenoliths and the dyke can hardly be questioned. The proportion of xenoliths of peridotite and pyroxenite to undoubted accidental xenoliths of Moine type bears no relationship to what one would expect were all the xenoliths accidental.

The coarse grain-size of the ultrabasic xenoliths and in particular the lamellar intergrowths of enstatite and augite in the pyroxenites

betokens slow crystallization of the type to be found in large intrusions of basic rock. The lamellar intergrowths and the two sets of augite lamellae in the enstatite resemble very closely those described, for instance, from the Bushveld Complex. One is left with the impression that monchiquite and ultrabasic xenoliths had the same parent magma, and that the xenoliths are the result of break-up of an ultrabasic fraction that formed from that magma before the dyke was intruded into its present position.

Other British occurrences of xenoliths of ultrabasic type have been described from the late Palaeozoic volcanic rocks of the Midland Valley of Scotland, from Caithness, and from Derbyshire, but appear to be unknown from the Tertiary igneous rocks of the Hebrides and Ireland. Examples which might be mentioned are the blocks of lherzolite, largely carbonated, in the Carboniferous Heads of Ayr vent (Tyrrell, 1918) ; a Permian dyke in the Sanquhar Coalfield (Simpson and Richey, 1936) seen in a mine, which contains abundant inclusions of altered peridotite ; and the intrusive analcite-basalt of Calton Hill, Derbyshire (Tomkeieff, 1928) which has abundant xenoliths of pyroxene-peridotite. Tomkeieff regards these xenoliths as due to the breakdown of a pre-existing rock rather than nodular segregations from the magma, as had been earlier believed. The most closely comparable example, however, appears to be the small intrusion of basalt described by Smellie (1915) from Bute, which contains fresh xenoliths of a variety of rock types, including dunite, various pyroxenites, gabbro, and anorthosite. Again, the probable Permian vent at Duncansby Ness in Caithness (Crampton and Carruthers, 1914) has associated dykes of monchiquite and nepheline-basalt and contains what are described as bombs of olivine rock.

There is a small proportion of xenoliths of other rock types in the Streap dyke. Quartz-bearing granulites can readily be matched amongst the rocks of the Moine Series, and these are undoubtedly accidental xenoliths. The pyroxene-gneiss obviously bears no genetic relationship to the dyke, and xenoliths of this rock-type must also be regarded as accidental xenoliths picked up by the dyke from its walls during intrusion. The pyroxene-gneisses are almost certainly metamorphic in origin, and in some respects, namely in their granulitic texture and in the presence of perthite, these rocks resemble members of the charnockite series. However, they are not true charnockites, for they do not contain quartz and not all contain hypersthene or clinohypersthene. In the hand-specimen also they show little resemblance to charnockites, and in fact more closely resemble banded anorthosites. Small areas of charnockitic rocks are known in the Lewisian of the north-west of Scotland (Sutton and Watson, 1951, p. 264) and the pyroxene-gneiss xenoliths in the Streap dyke may

possibly have been derived from a Lewisian-like complex lying below the Moine Series of the Glenfinnan area.

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Welded Tuffs in the Borrowdale Volcanic Series, English Lake District, with a Note on Similar Rocks in Wales

By R. L. Oliver

Abstract

Compact " streaky " rocks, from the Borrowdale Volcanic Series, English Lake District, are diagnosed in part as welded tuffs. Original glass is completely replaced, but similarity to welded tuffs elsewhere is shown by the dacitic-rhyodacitic composition, lack of bedding, and development of columnar jointing. The deposition of welded tuffs, as nuée ardentes, or incandescent tuff flows, is discussed.

Introduction

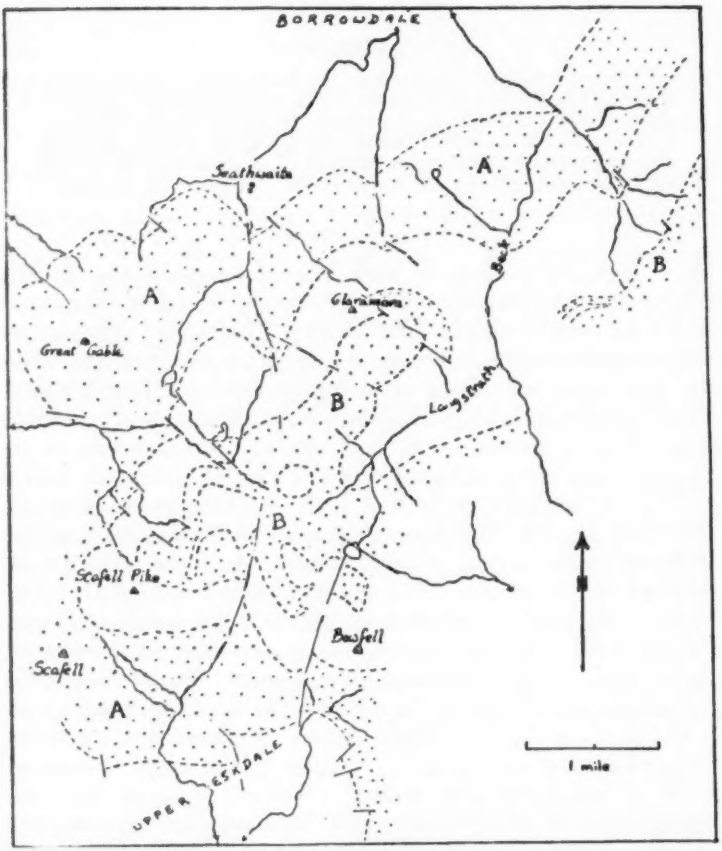
ALTHOUGH Iddings, in 1899, described some of the rhyolitic rocks of Yellowstone National Park as consisting of " fragments of pumice pressed together and welded ", widespread diagnosis of rocks as welded tuffs is a development of the last two decades. Mansfield and Ross (1935) define the welded tuffs of south-east Idaho as " tuffs in which the rock particles have retained their heat sufficiently after falling to become welded and solidified so that many of the resulting rocks simulated actual flows ". In New Zealand, welded tuffs, or " ignimbrites " (Marshall, 1932, 1935), the mode of deposition of which had for long been in question (see Grange, 1937), exhibit fluidal structures and are columnar jointed. Such characteristics are exhibited by the Brisbane Tuff (Richards and Bryan, 1934). Also similar to Marshall's ignimbrite is the Bishop Tuff of eastern California (Gilbert, 1938), with absence of bedding, distortion of vitric fragments due to weight of overlying material, and greater relative compaction and welding near the base of the deposit. The Wineglass Welded Tuff, of Crater Lake National Park, Oregon (Williams, 1942) contains " larger lumps of vitric glass . . . flattened by the weight of material above ". Ignimbrites with marked " fluidal " structure and rude prismatic jointing perpendicular to the " flow structure " are described by Osborne (1950) among the Kuttung Volcanics of New South Wales ; and Barksdale (1951), describing welded tuffs in the Lewis and Clark County, Montana, considers that the constituent fragments were plastic when deposited.

The Lake District Welded Tuff

(*a*) *Stratigraphical Relationships.*

Rocks diagnozed by the writer as welded tuffs, interbedded with rhyolite, rhyodacite, and dacite flows, and occasional layers of more normal pyroclastics, are mapped as a single unit (A, Text-fig. 1) and

termed the Airy's Bridge Group. The Group includes roughly Green's (1915) "streaky rocks", and corresponds to Marr's (1916, Text-fig. 5) Sty Head Garnetiferous Group. Other welded tuffs, here called the Lincomb Tarns Formation, which do not exhibit the



TEXT-FIG. 1.—Map of part of the English Lake District, showing distribution of formations with welded tuffs.

diagnostic characteristics of such rocks so commonly as those of the Airy's Bridge Group, outcrop at a higher stratigraphical level (B, Text-fig. 1). Both units are part of the Borrowdale Volcanic Series, a sequence of basaltic, andesitic, and rhyolitic flows and tuffs, of Llandeilo age (Marr, 1916, p. 17), and are thus older than any recorded occurrences of welded tuffs known to the writer.

The calculated thickness of the Borrowdale volcanics within the

mapped area (Text-fig. 1) is approximately 10,000 feet, of which the Airy's Bridge Group comprises a maximum of 2,850 feet and the Lincomb Tarns Formation 800 feet. The latter cannot justifiably be subdivided, but within the Airy's Bridge Group can be recognized :

	Ft.
1. Rhyolite and rhyodacite flows, flow-brecciated in part	0–200
2. Rather coarse tuff, not strongly welded	0–150
3. Flows and welded tuff of rhyodacitic-dacitic composition	2,500
4. Welded dacitic tuff, with some normal bedded tuff	0–760

(*b*) *Field Characteristics.*

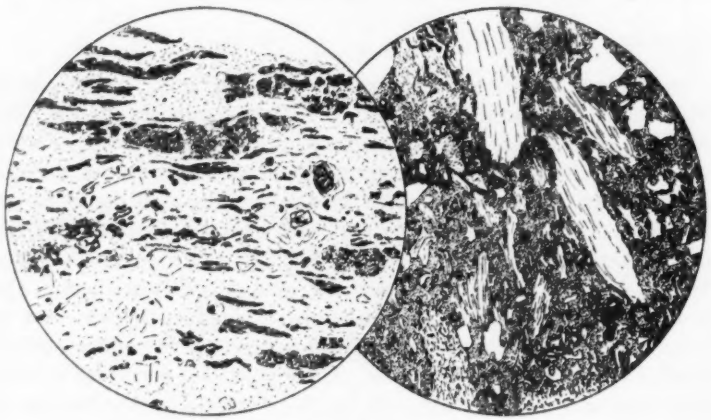
In general, the Lake District welded tuffs are flinty, and unbedded. Their chief diagnostic feature, however, is the appearance, on most surfaces, of lenticular streaks, orientated parallel to each other and to the plane of deposition. In some cases, the streaks are finer and less discrete than are those illustrated, and closely resemble the true flow lineation of lavas. This apparent flow structure, depicted by the parallelism of the thinner streaks, is in keeping with the resemblance to fluidity noted above as characterizing welded tuffs elsewhere. The lenticular patches are the cross sections of roughly-shaped discs or plates, the broad dimensions of which are visible only on a surface parallel to the plane of deposition. These plates are presumed to have been originally plastic fragments, as, for example, in the Wineglass Tuffs, " flattened by the weight of material above." In the Lake District rocks, the streaks, which are characteristically weathered below the surface of the surrounding rock, attain a length of six inches, though they are usually smaller, like those of the New Zealand ignimbrite (variety " owharoite " [1]). Williams (1942) describes black strings of obsidian up to a foot long in the Wineglass tuffs.

As elsewhere, the resemblance of some of these streaks to true flow structure in the Lake District rocks renders distinction of welded tuffs from lavas extremely difficult. The difficulty is noted by Walker (1904, p. 94), and is alluded to by Marr (1916, p. 23) when he refers to the streaky structure " occurring in both lavas and fragmental rocks ". Green (1919, p. 163), in discussing the vulcanicity of the Lake District, notes that " there is complete gradation from rhyolites with rounded or subangular enclosures, having a disquieting resemblance to a tuff when weathered, to a well-banded rock ". It is of interest that the similarity (indicated by the written account only)

[1] The " owharoite " is in reality an example of the " wilsonite " type of ignimbrite, from Owharoa, North Island, New Zealand. " Wilsonite " is one of the several types of ignimbrite distinguished by Marshall (1935, pp. 332–8).

of these Lake District rocks, described by Green, to the New Zealand ignimbrites was noted by Marshall (1935, p. 356). That definite flow rocks do occur in the Airy's Bridge Group has already been noted. Where these are streaky, the streaks, in addition to being generally finer and more continuous, tend to be pencil-shaped rather than disc-shaped. However, similarity of the welded tuffs and flow rocks is such as to make separation impossible in mapping much of the Airy's Bridge Group.

Resemblance to the streaky structure of the welded tuffs results also



A B

TEXT-FIG. 2.—A. Welded tuff from Adam-a-Crag, upper Eskdale (Min. Pet. Dept., Cambridge, Slide No. 77019). Lenticular "streaks" composed essentially of a quartz-chlorite granular mosaic, partially epidotized albite crystal fragments, iron ore, in a fine-grained felsitic matrix. × 22.

B. Ignimbrite (variety "owharoite") from Owharoa, North Island, New Zealand (Min. Pet. Dept., Cambridge, Slide No. 57258/a). Flattened glass lapilli and scattered oligoclase and quartz crystal fragments in a matrix of smaller shards and glass dust. × 22.

from shearing. Streaks of this kind, however, are, in most cases, oblique to the plane of deposition, and can usually be distinguished.

Welded tuffs of the Airy's Bridge Group are columnar jointed in places. A good example may be seen in a small stream approximately 750 yards south-east of Seathwaite-in-Borrowdale. Well-developed columnar jointing is developed also at other localities in particularly hard and compact, flow-like welded tuff at the base of Division 3 of the Airy's Bridge Group.

Welded tuffs in the Borrowdale Series are apparently not restricted to the area mapped by the writer. Thus, roughly parallel, lenticular

depressions, occurring on the weathered surface of some of the Coarse Tuffs (= Lincomb Tarns Formation ?) in the south-eastern Lake District are mentioned by Mitchell (1929, p. 23) and he suggests that they represent fragments which have been flattened by the weight of overlying material. In 1934, Mitchell records "streaky tuff" from Sleddale Fell, also in the south-east; and Hartley (1942, p. 143), describing some of the rhyolites in the Thirlmere-Helvellyn area, mentions the occurrence of a "'streaky' texture, as distinct from normal flow structure", and says that a similar texture occurs in some of the Langdale rhyolites.

(*c*) *Mineralogy and Chemistry.*

In a typical rock (Text-fig. 2*a*), from near Adam-a-Crag, upper Eskdale, the "streaks", which are most apparent, consist of a granular intergrowth of chlorite and an approximately equal amount of felsitic material which is mainly quartz with some alkali feldspar. Scattered albite "phenocrysts" are fairly numerous. A little epidote is associated with the plagioclase and, contained also in the section, though not pictured in text-fig. 2*a*, are one or two chlorite pseudomorphs after pyroxene (?). The felsitic mesostasis consists essentially of fine-grained potash feldspar and quartz, with scattered granules of partially leucoxenised iron-ore, very minor apatite, and a few wispy strings of sericitic material developed under stress.

The welded tuff from Slate Crag, Bowfell, and which is very similar to that just described, has the composition (Table 1) of a typical rhyodacite. Flow rocks and other welded tuffs from terrain which has been mapped as the Airy's Bridge Group—analysed by the writer, by Ward (1875), and by Walker (1904)—range in silica content from

TABLE 1.—CHEMICAL ANALYSIS OF WELDED TUFF FROM SLATE CRAG, BOWFELL, ENGLISH LAKE DISTRICT. Analyst: R. L. Oliver.

SiO_2	67·05
Al_2O_3	16·03
TiO_2	0·28
Fe_2O_3	0·73
FeO	4·02
MnO	tr.
MgO	0·44
CaO	1·97
Na_2O	3·59
K_2O	4·01
P_2O_5	0·29
CO_2	0·05
H_2O+	1·37
H_2O-	0·16
	99·99

59·04 per cent to 69·67 per cent, and a welded tuff from the Lincomb Tarns Formation has 63·08 per cent SiO_2. Most welded tuffs from elsewhere are rhyolitic, with dacitic and latitic compositions also represented; thus constituting a group with a limited compositional range to which the Lake District rocks satisfactorily belong.

Consideration of the New Zealand ignimbrite (variety owharoite) as seen in thin section (Text-fig. 2*b*), shows that, apart from a few crystal grains of oligoclase and quartz, the rock consists entirely of glass fragments—a mixture of dust, small shards, and larger pieces in distinct parallel arrangement; the latter have been compressed perpendicular to the plane of deposition, according to Marshall (1935), and as noted above. Apparently, almost identical are the Wineglass Tuffs, of Crater Lake National Park, Oregon, with a matrix composed of densely packed shards and interstitial glass dust, and in which " the obsidian streaks represent deformed clots of magma . . ." (Williams, 1942).

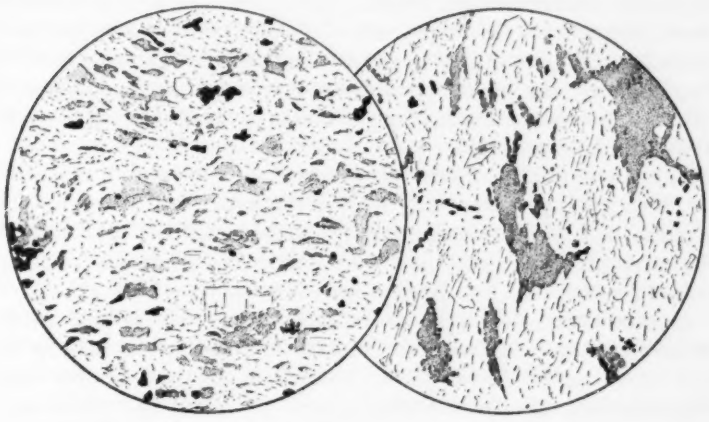
The question now arises as to how strongly the macroscopic resemblance of the Lake District and New Zealand rocks is substantiated by the thin section comparison. Although in the Lake District example (Text-fig. 2*a*), all indication of the former presence of angular glass shards has completely disappeared, there is little doubt that the felsitic groundmass is a devitrification product as in other (probably flow) rocks of the Airy's Bridge Group which have a similar groundmass cut by perlitic cracks. The derivation of the chloritic streaks from flattened vitric fragments is more uncertain though D. Williams (1930) describes broken glass shards devitrified to a quartz-chlorite mosaic in vitric tuff of Ordovician age from north Wales.

Welded Tuffs in Wales

H. Williams (1927, p. 373) refers to the impossibility, concerning some of the Ordovician Snowdonian volcanics, of deciding whether the rocks are tuffs or lavas. Later (personal communication, 1952) he states that welded tuffs are undoubtedly present in north Wales. A specimen in the collection, Mineralogy and Petrology Department, Cambridge, labelled Rhyolite Tuff, from near Glyder, Carnarvonshire, is flinty in hand specimen, and exhibits lenticular chloritic clots similar to those of the Lake District rocks under consideration. In thin section (Text-fig. 3*b*), may be seen the chloritized fragments, perhaps modified a little in shape by shear, with a fine-grained mesostasis of alkali feldspar, quartz, chloritic and sericitic material, representing devitrified glass dust and small shards.

In Pembrokeshire, south-west Wales, what were thought to be lavas, also of Ordovician age, exhibiting " flow structure ", were described

by Reed as long ago as 1895. One of these rocks, from two miles east of Fishguard, Pembrokeshire, shows, under the microscope (Text-fig. 3*a*), the outline of former glass shards, now devitrified to chlorite, and which appear to have been drawn out by flow rather than compressed vertically. Alkali feldspar, quartz, and chlorite granules—also probably the product of devitrification—constitute the intervening ground.



TEXT-FIG. 3.—A. Welded tuff, from 2 miles east of Fishguard, Pembrokeshire (Min. Pet. Dept., Cambridge, Slide No. 34389). Chloritized shards, quartz, and sericitized acid plagioclase crystal fragments, shredded granules of iron ore, in a felsitic base. × 44.

B. Welded tuff, "near Glyder, Carnarvonshire" (Min. Pet. Dept., Cambridge, Slide No. 3987). Chloritized lapilli, modified in shape by shear, in a felsitic base. × 22.

DEPOSITION OF WELDED TUFFS

In view of the combination of pyroclastic and flow characteristics possessed by welded tuffs, their lack of bedding, and the evident initial heat of the constituent glass fragments, it is not surprising that welded tuffs are considered by most workers to have originated as nuée ardentes, or incandescent tuff flows, such as have been observed erupting from Mt. Pelée, Martinique (Lacroix, 1904 ; Perrett, 1935), and are considered to be the origin of deposits in the Valley of Ten Thousand Smokes, near Mt. Katmai, Alaska formed earlier this century (Fenner, 1923 ; 1937). Some observers (e.g. Gilbert, 1938 ; Barksdale, 1951) prefer comparison of welded tuff only with the deposits in the Valley of Ten Thousand Smokes ; and, indeed, there would seem to be little, if any, similarity between the unconsolidated material, erupted from Mt. Pelée, and the typically compact, welded

tuff or ignimbrite (see also MacGregor, 1946). H. Williams (1927) favours an origin for some of the non-stratified fragmental rocks of Snowdonia similar to that of the " great sand flow " in the Valley of Ten Thousand Smokes rather than to that of the Mt. Pelée nuée ardente material as suggested earlier by Dakyns and Greenly (1905).

Even the incandescent sand flow of the Valley of Ten Thousand Smokes cannot in total be classified as a welded tuff, for much of the deposit is unconsolidated (Fenner, 1923 ; 1937). Also, some of the more compact rock has been " welded " as the result of pneumatolytic crystallization (Fenner, 1948), and not because of the plasticity of the fragments and the weight of overlying material or to the indigenous devitrification of the shards, as is the case in a true welded tuff (Fenner 1948 ; Barksdale, 1951).

In mode of origin, however, there would seem to be no justification in disassociating welded tuffs from incandescent tuff flows on the grounds that deposits resulting from the latter are not consolidated for some recognized welded tuffs are part only of a deposit which is essentially unwelded (e.g. Bishop Tuff—see Gilbert, 1938).

In postulating the origin, as an incandescent tuff flow, of those welded tuffs like the owharoite variety of New Zealand ignimbrite, and the Wineglass Tuff of Crater Lake National Park, Oregon (and also, therefore, like those of the English Lake District), special consideration should perhaps be given to the significance of the constituent once-plastic lumps of glass which have been compressed vertically. That the compression operated on a stationary mass seems probable, for the flattened fragments are, as noted in the description of the Lake District welded tuffs, roughly disc-shaped, with no tendency to the ellipsoidal form which would result from the influence of flow. It must be assumed, therefore, that the mass had flowed to a standstill before distortion or compression of the fragments had taken place. It must be assumed, also, that the mobility of the mass of " highly inflated pumice " (of which, on analogy with Katmai, the material may be assumed to consist) was, on extrusion, virtually that of a liquid, and that the mass ceased to flow only when horizontality had been attained ; for cessation of flow on much more than a slight gradient could have resulted only by the development of viscosity due to cooling, a viscosity which, surely, would have preserved the signs of final flow. Confirmation of extreme mobility of welded tuff material is afforded by the plateau-like surfaces of many of the world's welded tuff deposits (see Cotton, 1944, pp. 206–208).

Welded tuffs in general, therefore, may be considered to have originated as the whole or part of nuée ardentes or incandescent tuff flows, though, as already noted, welding is not an essential feature of such tuff flows, many of which have now been described as being

completely unwelded and incoherent. Material, ejected as a tuff flow or glowing cloud, by the main explosion of Mt. Mazama, in the Crater Lake National Park, Oregon, is virtually unwelded (Williams, 1942), though it overlies directly the thoroughly welded Wineglass Tuff. Williams (1942), considers that variations in the degree of welding are due to the nature of the underlying topography, and to differences in temperature, gas content, and thickness of the ejected material. Presence or absence, therefore, of these properties in correct proportion, perhaps determines whether or not welded tuff is formed from " the chaotic unbedded deposits left by glowing clouds " which " have been identified in scores of volcanic regions and may now be considered characteristic of gas-rich intermediate and acid lavas " (Williams, 1941, p. 379 ; and see also Cotton, 1944, p. 205).

Conclusion

Rocks which include Walker's (1904) and Green's (1915) " streaky rocks ", and which were mapped by Marr (1916, Fig. 5) as the Sty Head Garnetiferous Group, have been remapped as the Airy's Bridge Group (A, Text-fig. 1), and are considered to consist largely of welded tuff (= ignimbrite). More welded tuffs, mapped as the Lincomb Tarns Formation (B, Text-fig. 1) occur at a higher stratigraphical level.

As noticed by Marr (1916, p. 23), the characteristic streaks occur " in both lavas and fragmental rocks " and, in many cases, distinction of the two rock types is extremely difficult. The streaks are a manifestation of viscous flow in the lavas ; in the welded tuffs they represent former glass lapilli, plastic when deposited, and flattened by the weight of the overlying material, as in similar tuffs elsewhere.

Complete devitrification of the Lake District welded tuffs has converted the lapilli into an essentially quartz-chloritic mosaic, and has replaced the presumed (on analogy with the New Zealand ignimbrite) matrix of shards and glass dust by a mesostasis of fine-grained potash feldspar and quartz, a little sericitic material, and ragged grains of partially leucoxenised iron ore.

Other diagnostic characteristics of welded tuffs, which demonstrate their resemblance to flow rocks, are compactness, absence of bedding, and the development of columnar jointing, all of which features are illustrated by the Lake District welded rocks.

What appear to be welded tuffs, somewhat similar to those in the Lake District, outcrop also in north Wales. Resemblance to true flow in an example from Pembrokeshire is due to the attenuation of once plastic shards rather than to the parallelism of vertically compressed lapilli.

Welded tuffs are considered by most workers to have originated as nuée ardentes, or incandescent tuff flows, and it is probable that those

of the English Lake District have originated in a similar manner.

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NOTICE

ZOOLOGICAL NOMENCLATURE

NOTICE is hereby given that the possible use by the International Commission on Zoological Nomenclature of its Plenary Powers is involved in applications relating to the undermentioned names included in Part 9 of Volume 9 of the *Bulletin of Zoological Nomenclature*, published on 22nd October, 1954:—

(1) Names (generic and specific) given to aptychi of Ammonites, proposed suppression of (pp. 266–9) (File Z.N. (S.) 858).

(2) *Cheloniceras* Hyatt, 1903 (Class Cephalopoda, Order Ammonoidea), proposed designation of type species for, in harmony with accustomed nomenclatorial usage (pp. 278–280) (File Z.N. (S.) 703).

(3) *minimus* Miller (J. S.), 1826, as published in the combination *Belemnites minimus* (Class Cephalopoda, Order Dibranchia), proposed validation of (pp. 284–5) (File Z.N.(S.) 823).

In addition, the above Part contains a proposal for the issue of a *Declaration* banning names based upon the aptychi of ammonites (File Z.N.(S.) 589).

The Plutonic History of the Aswan Area, Egypt

By A. R. Gindy

Abstract

The coarse porphyritic Aswan Granite used in many ancient Egyptian monuments has been held to intrude a tract of gneisses and schists. It is suggested in this paper that the granitic body was formed during a phase of granitization which followed the folding and metamorphism responsible for producing the gneisses and schists. A part of the granite retains mimetic structures inherited from the pre-existing rocks, but the southern part became mobile and developed a new structure. The intrusion of a magmatic granite in dykes and sheets followed the phase of granitization.

Introduction

THE Aswan area in the Nile Valley has been famous for more than five thousand years as the gateway to Nubia and the Sudan, and as the source of many beautiful rocks used for monumental works. The region is of special interest to geologists in Egypt as it was until recently almost the only easily accessible area of metamorphic and igneous rocks, and an extensive literature on it has therefore grown up.

The present writer recently completed a detailed geological survey on a scale of 1 : 10,000 of an area in the Nile Valley roughly bounded by longitudes 32° 51′ E. and 32° 56′ E. and latitudes 24° 6′ N. and 23° 59′ N. (see text-fig. 1). The field and laboratory data obtained during this study have led to new conclusions concerning the plutonic history of the area. The oldest rocks appear to represent a thick sedimentary series, with intercalated basic igneous intrusions, which was intensely folded and sheared and at the same time injected by acid igneous fluids. After this diastrophism, granitization of the country-rocks by the acid fluids took place, and a body of coarse porphyritic granite was formed. This granite, which occupies the central part of the area and supplies the monumental stone of Aswan, is mainly of migmatitic origin although large parts of it have moved since their formation. At later stages a finer-grained magmatic granite was intruded in the form of sheets and dykes, local belts of strong shearing were developed and, finally, the intrusion of a number of varieties of lamprophyric and basic rocks took place.

All the igneous and metamorphic rocks, with the exception of some of the youngest minor intrusions, were peneplaned and covered unconformably by almost horizontal Nubian Sandstone. This formation is of Cretaceous age, and recent writers are agreed that the underlying metamorphic rocks cannot be younger than Carboniferous and may be Precambrian.

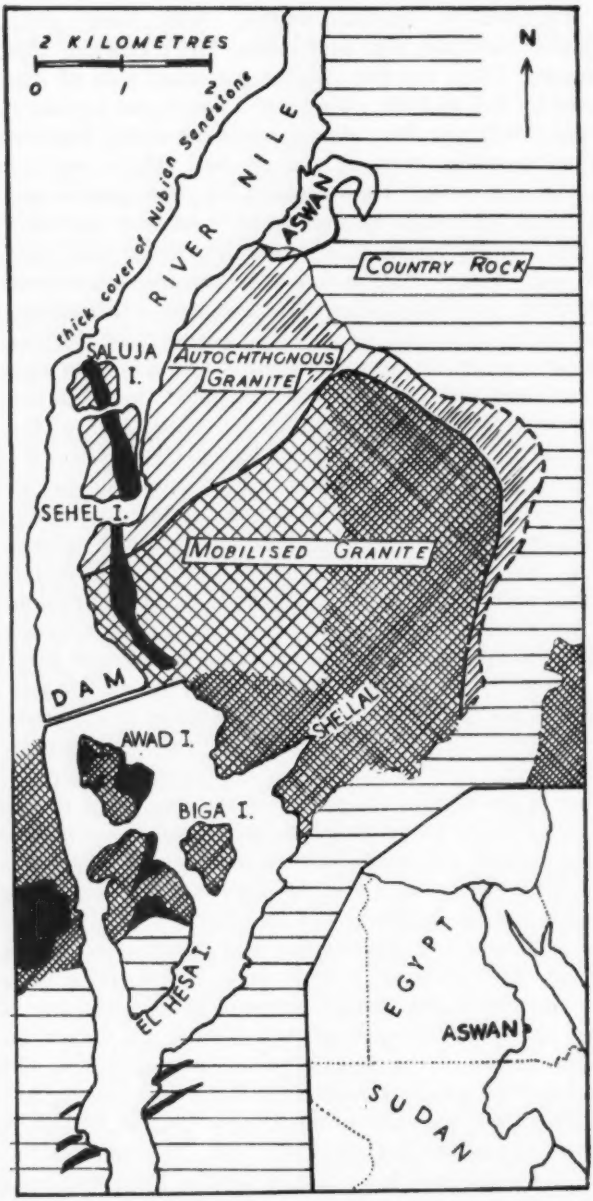
The conclusions outlined above differ in many respects from the

views expressed by previous writers. After many early petrographical studies (see, e.g., Hawkshaw, 1867; Dawson, 1884, 1886; and Bonney, 1886), the first detailed geological map of Aswan was published by Ball in 1907. A map of a wider area is given by Little and Attia (1943) who have also reviewed the earlier literature. Ball recognized an older series of gneisses and schists and a younger intrusive granite series. He considered the older gneisses and schists to be highly deformed igneous rocks. Andrew, however (1934), recognized among these older rocks psammitic and semi-pelitic schists and gneisses interbanded with hornblende-biotite-andesine-quartz schists, which sometimes showed traces of a sedimentary origin, and sometimes appeared to represent altered basic igneous rocks. The present writer considers the gneisses and schists to represent a metamorphosed series of sediments and minor basic intrusions.

The granitic body which occupies the central part of the region described in this paper was considered by Ball (1907), Barthoux (1922), and Andrew (1934) to be an igneous intrusion. Ball and Barthoux recognized within it a number of dioritic, anorthositic, syenitic, and granodioritic varieties which they regarded as distinct igneous types, although they noted many transitions between the different varieties. Barthoux assigned these types to a definite series of intrusions. Ball noted that dioritic patches in the granite were often associated with hornblende-schists which he regarded as the products of deformation of the diorites. In the present writer's view, the dioritic and granodioritic varieties appear to be intermediate stages in the granitization of the country rocks, the schists associated with such varieties being the parents and not the derivatives of the massive rocks. Some confusion in determining the history of the area seems to have been caused by failure to differentiate between the early foliation developed along the bedding and later planar structures imposed by shearing.

The rocks are described in the following pages as nearly as possible in chronological order, beginning with the country rocks and progressing to their deformation, metamorphism, and granitization; an account of the coarse Aswan Granite, which is the final product of granitization, follows, and finally the younger intrusive granite, the late belts of shearing, and the various late minor intrusions are dealt with.

The distribution of the more important rock groups is shown in text-fig. 1 and the structural features of the area in text-fig. 2. It can be seen from these maps that a broad belt of gneisses and schists runs north and south on the east side of the Nile. The main granite of the area shown on text-fig. 1 has a roughly circular form, with a diameter of 4–5 kms. and occupies the central part of the area. On the west side of the Nile both granite and metamorphic rocks are largely covered by the Nubian Sandstone.



TEXT-FIG. 1.—Geological map of the Aswan district. Diagonal lines: autochthonous granite. Cross-hatching: mobilized granite. Black: basic rocks. Close hatching indicates increasing mobilization of granite.

TEXT-FIG. 2.—Structural map of the Aswan District.

The Country Rocks

(*a*) *Original nature.*

The country-rocks appear to have consisted originally of a thick, possibly geosynclinal, series of alternating semipelitic sediments with subordinate psammitic bands. No carbonate rocks have been found within the area shown on text-fig. 1. Beds of basic tuffs and ashes (perhaps represented by the sedimentary hornblende-schists described by Andrew, 1934) may be present, but are now more or less indistinguishable from rocks derived from minor basic intrusions. These latter bodies form small sills and dykes which were intruded into the sediments before their metamorphism. They appear to have been gabbroic in composition and to have consisted almost entirely of pyroxene and basic plagioclase with subordinate ilmenite. The smaller intrusions, and the margins of some of the larger bodies, are very fine-grained, but the central parts of the larger intrusions may reach a grain-size of 5 mm. Original sub-ophitic and doleritic textures can sometimes be found.

(*b*) *Orogenic deformation, accompanied by acid igneous intrusion.*

After the intrusion of the basic rocks into sediments, the whole series was subjected to strong diastrophism during which the rocks were thrown into intricate folds of every dimension and most of their foliate structures were impressed upon them. It will be seen from text-fig. 2 that in the north and east of the area the strike is north-north-east but that it swings round into a south-westerly direction to the south of the granite; this diversion of strike is probably due to the effects of late belts of shearing. The formation of a marked lineation trending roughly parallel to the strike accompanied the folding.

During the diastrophism, acid pegmatitic material was intruded into the deformed rocks in small sheets, sills, and veins a few feet to a few inches in thickness. That these bodies were contemporary with the deformation is shown by the fact that they are folded along with the surrounding rocks and are sometimes drawn into lenses or boudins between which the country rocks flowed. Throughout the altered sediments it is possible to recognize in addition to these early deformed intrusions, later bodies of almost the same material which are unsheared and which form tortuous veins crossing the earlier pegmatites. These post-diastrophic veins become diffuse, mixing with the country rock and helping to produce the effects of regional metamorphism and granitization which are described in the next section.

The presence of early discrete acid intrusions indicates that at the start of regional metamorphism and granitization a genuinely liquid

material was available. It also establishes the composition of this material. In the oldest and most strongly sheared intrusions oligoclase is often the predominant feldspar, but in the later veins the two feldspars are present in nearly equal amount and, with quartz, make up almost the whole rock. Ferromagnesian minerals form at most 2 per cent of the rocks and are often completely absent. When present, they include biotite, hornblende, garnet, sphene, apatite, and opaque minerals, all of which may be due to contamination by the country rock. The acid fluids must thus have been composed almost entirely of silica, alumina, alkalies, and water vapour. They were at first sodic and later sodi-potassic.

Progressive Metamorphism and Granitization

The continued introduction of acid material in a diffuse manner led in the initial stages to the development of oligoclase and then of potash feldspar porphyroblasts in the banded sediments and associated basic intrusions, and then to the gradual elimination of pre-existing structures culminating in the passage of the country-rocks over wide areas into coarse porphyroblastic granite or granodiorite. The earlier stages in this process are to be seen in the country rocks proper, outside the boundary of the granite mass (text-fig. 1) while the intermediate and final stages are confined to a zone within a few hundred yards of the granite boundary, and to relics of country rock in the granite itself. Within the country rocks proper, oligoclase porphyroblasts do not exceed 5 mm. in length, but at the granite margin they may reach a centimetre or more.

Within the area shown on text-fig. 1 there is a progressive increase in the degree of metamorphism and granitization of the country rocks from north to south. North of the granite, the rocks are relatively fine-grained and oligoclasic, except along the border of the granite; further south they are gneissose, and are richer in potash feldspar. There are, however, many local variations in the degree of alteration which are probably due to a patchy distribution of the granitizing fluids.

The type of contact shown by the granite also varies from north to south. In the north it tends to be highly irregular and transitional and shows the familiar features of granitized terranes (i.e. the development of granitic patches and *dents de cheval* in the bosom of the country rock and the preservation of structurally undisturbed relics of country rock in the granite). South of Shellal the granite contact appears to be sharp and intrusive, although the country rocks are themselves in a more advanced stage of alteration than those further north.

(*a*) *Metamorphism and granitization of semi-pelitic, pelitic, and*

psammitic rocks.—The earliest recognizable stage in the alteration of pelitic and semi-pelitic rocks is represented by dark or grey schists containing oligoclase (which makes up to 55 per cent of the rock), quartz, biotite, and relatively small quantities of potash feldspar. The average grain-size is usually well below 0·3 mm.

In the next stage of alteration the oligoclase grains increase in relative size and become rounded or subhedral, replacing the mosaic along their borders in a sinuous manner and enclosing occasional grains of the matrix minerals in a manner such as has been described by Read from Cromar, Aberdeenshire (1927) and from Central Sutherland (1931). As growth of the porphyroblasts continues, lamellar twinning and irregular oscillatory zoning become more prominent and adjacent crystals may unite to form a single larger porphyroblast of clumsy construction. The general grain-size of the rock increases and biotite and quartz recrystallize in the spaces between the feldspars. Potash feldspar appears later than plagioclase, and forms hair-like intergranular films which continue to grow porphyroblastically, replacing the matrix minerals and the plagioclase porphyroblasts. The latter are either replaced in an irregular manner or are surrounded by a mantle of potash feldspar. Ultimately the potash feldspar crystals become larger and more robust than the plagioclase. At the same time, fantastic growths of myrmekite appear, in amounts apparently inversely proportional to the amount of perthite in the potash feldspar.

During the early stages of granitization, many of the feldspar porphyroblasts lie parallel to the foliation of the rocks; their orientation does not appear to be due to contemporaneous or later deformation, and was probably a result of growth along the planes of least resistance where the supply of introduced material was most abundant. In advanced stages, the orientation may be lost, but a weak relict foliation or lineation generally persists and can usually be recognized in the granite itself. When granitization is complete, the original fine-grained matrix is almost totally recrystallized and a coarse granitic rock is produced; occasional relics of the fine-grained texture can be seen in thin sections of such granitic rocks.

At intermediate stages in the process of granitization, sillimanite fibres develop, apparently by replacement of biotite. Large garnets, poikiloblastically enclosing grains of the matrix, are locally numerous. Accessory apatite, zircon, sphene, ilmenite, magnetite, epidote, and allanite increase appreciably during granitization, and rare hornblende appears in the country-rock relics within the granite.

The few psammitic bands in the country rock follow a course of alteration similar to that of the pelites and semi-pelites but, judging by the number and state of crystallization of their relics in the granite, their alteration lags behind that of the other rocks. The least altered

varieties show a fine, granular, almost felsitic, mosaic in which potash feldspar may sometimes be present. During granitization the grain-size increases and feldspars, biotite, and the accessory minerals are introduced.

(*b*) *Progressive metamorphism of basic rocks.*—In the earliest stages of alteration, some of the igneous textures and minerals of the small basic intrusions into the sedimentary series are preserved, although the pyroxene is usually partly replaced by uralite and the labradorite is clouded. This stage of metamorphism can be found all through the country rocks north of the latitude of the dam.

As alteration proceeds, the texture becomes granular, plagioclase recrystallizes to clear labradorite, and pyroxene is completely replaced by pale granular hornblende. The anorthosites mentioned by Barthoux represent the few cases where this pale hornblende falls to less than 14 per cent of the rock. Thereafter the plagioclase becomes more acidic, reaching the composition of andesine or basic oligoclase, the hornblende grows darker, quartz grains appear, and the original ilmenite develops rims of botryoidal sphene. Rocks in this state of metamorphism probably represent the diorites of previous writers.

With more advanced alteration, plagioclase develops into clumsy and irregularly zoned porphyroblasts. Potassic ingredients first appear with the conversion of hornblende into biotite; thereafter, films of potash feldspar replace the other minerals, especially the plagioclase, and ultimately form large porphyroblasts. Sphene, apatite, and allanite all increase in amount.

The rocks at this stage present a peculiar appearance, with large red or white porphyroblasts set in a darker and finer-grained matrix made up of plagioclase, hornblende, biotite, and quartz. Rocks of this type were termed syenite by Ball (1907), but Barthoux (1922, p. 135) pointed out that they could be more correctly described as granodiorites. Both writers regarded them as distinct igneous types. They are, however, found in the field to form crude zones around cores of less altered basic rock; on Sehel Island, granodioritic rocks can locally be seen to contain dark angular relics of basic rock which give them an agmatitic appearance. The diorites and granodiorites are therefore regarded by the author as intermediate stages in the granitization of basic rocks. With further alteration, quartz and potash feldspar increase, hornblende and biotite diminish, and the rocks pass into the characteristic coarse Aswan granite.

The great majority of the country-rock relicts in the granite are derived from the minor basic intrusions, and this fact illustrates the lag in the rate of granitization of these rocks in comparison with the rate of alteration of pelitic and semi-pelitic rocks. In the northern part of the granite some of the old basic intrusions are preserved

practically intact and have almost the appearance of cutting the granite.

The changes described above indicate that the basic rocks have suffered an increase in silica and alkalies, substances which characterize the acid intrusions that invade the country rock and frequently form discrete or cloudy veins in basic rock relics within the granite. Lime is the principal ingredient lost during granitization of the basic rocks. Some lime is absorbed by the formation of sphene from ilmenite and by the development of apatite. Some is probably fixed in the outer zones of reaction around the hornblendic relicts, where potash feldspar porphyroblasts frequently become surrounded by rapakivi mantles of plagioclase. Similar mantles sometimes enclose potash feldspars in the granitic rocks derived from pelitic or semi-pelitic parents near the margins of basic inclusions and illustrate the basifying effect of the out-going lime. There is no evidence to indicate the existence of large-scale basic fronts.

The Coarse Aswan Granite

The Aswan Granite, which is the final product of the processes of metamorphism and granitization described in the last section, is the famous rock from which many ancient Egyptian monuments (including the obelisk on the Thames Embankment, London, and most of the colossi in the British Museum) were made. This rock was originally called syenite by Pliny (Syene is the ancient name of Aswan) but recent practice has so changed the meaning of the word that it is no longer applicable to the rocks of Aswan. In modern terminology the dominant rock of the Aswan body is a granite.

In most parts of its outcrop the Aswan Granite presents a heterogeneous and highly migmatitic appearance and contains abundant small relics of the country rocks, especially of the basic rocks. The granite itself is rarely free from planar or linear structures produced by a rude alignment of the large feldspars. In some localities these structures represent the relict foliation and lineation of the country rock; elsewhere they are flow structures produced by movements of the granite. Shearing stresses seem to have been absent during the formation of the granite although a phase of shearing took place later.

The typical Aswan Granite is made up of large euhedral crystals of potash feldspar, often more than an inch long, crowded together in a darker matrix of quartz, oligoclase, hornblende, biotite, sphene, and apatite. The modal percentage of the dark minerals seldom falls below 5 per cent but hornblende may be scarce or absent.

Although the granite is considered to be of granitization origin and to have developed in place from the country rock, there are strong

indications that the south and south-east part of the body became mobile and flowed. The approximate distribution of autochthonous and mobilized granite is shown in text-fig. 1, but the line of demarcation is purely arbitrary and the two varieties show insensible gradations into one another. Moreover, the autochthonous part of the body contains numerous isolated pockets of homogeneous granite very like the mobilized granite of the southern region. In order to bring out the contrast between the autochthonous and the mobilized granite the characters indicating the origin of each are set out separately below.

The autochthonous granite : evidence of granitization origin.

(i) The transitional passage of the country rock into granite and of relict inclusions into granite has already been described. Large isolated feldspars, sometimes with rapakivi mantles, are developed in the midst of the country rock and granitic patches can be found within a few hundred yards of the granite boundary.

(ii) Replacement textures are observable under the microscope and have already been described.

(iii) The linear structures imposed upon the country rocks before granitization are preserved in their original orientation in completely isolated relict inclusions in the granite. A weak relict lineation is often preserved in the granite itself and is parallel with that in the country rocks, even where it crosses the boundary of the granite at a considerable angle (text-fig. 2). There has therefore been no disturbance of the structure during the formation of the granite.

(iv) The lag in the rate of granitization of the altered basic rocks with respect to the pelites and semi-pelites associated with them has resulted in the preservation of the former when the latter were completely granitized. The relics of basic rock remain unsupported in the granite, retaining their original dyke or sheet-like form for several yards and might at first sight be taken for post-granitic dykes were it not for the fact that they are traversed by feldspathic veins and can be seen (e.g., on the Island of Sehel and on the shore opposite the island) to pass into angular inclusions in the granite.

The mobilized granite : evidence of rheomorphism.

(i) Although, as has already been mentioned, the country rocks in the southern part of the area are more highly migmatitic than those in the north, their contact with the granite south and east of Shellal is rather sharp, as are the contacts of the inclusions in this region. This is probably due to the motion of the softened granite which dragged away any transitional material originally formed along the contact or around the inclusions.

(ii) The granite is more homogeneous in the east and south-east than it is elsewhere, and the majority of the granite quarries are situated in this region. The only inclusions remaining are of basic rock and these are less abundant than they are in the north.

(iii) In the south-eastern part of the granite inclusions of basic rock occasionally contain a light green pyroxene newly formed at the expense of hornblende, which is never found in the northern part of the region. The presence of this mineral probably indicates a higher grade of metamorphism.

(iv) What may be taken as an indication that parts of the mass, even in the northern region, became fluid, is the occurrence in a few localities of oligoclase porphyroblasts with a potash feldspar mantle side by side with rapakivi feldspars with a potash feldspar core and a plagioclase mantle. The latter, which occur in the vicinity of the basic inclusions, probably developed in the solid, whereas the former may have crystallized direct from the liquid portions of the rock.

(v) In the southern part of the granite the trend of lineations in the inclusions differs from the regional trend and varies from one inclusion to the next. Some inclusions show an almost horizontal foliation, although the dips of the country rocks are always steep. The granite itself may show planar flow structures whose trend is independent of the strike of the country rocks (text-fig. 2).

From the evidence quoted above it is more reasonable to regard the south-eastern part of the Aswan Granite as a very highly granitized region in which the rocks reached a state of mobility and were capable of flowing, than as an intrusive body which developed a migmatite complex along its northern boundary. It should be noted that in addition to the silica, alumina, and alkalies brought in by the original acid intrusions, an important part of these ingredients, together with lime, magnesia, iron, titania, etc., was provided on the spot by the country rocks themselves. The basic ingredients which were originally concentrated in the minor basic intrusions were finally distributed uniformly through the granite. In areas where the chemical changes came nearest to completion, and where the rocks were probably enriched in water and fugitive constituents, the granitic rocks reached a mobile state. As is to be expected in a granitization granite, modal and chemical analyses are variable, but the last two analyses to be published (Barthoux, 1922, and Harwood, in Hume, Harwood, and Theobald, 1935) do not differ markedly in the percentages of the major constituents.

The New Granite

After the diastrophism, metamorphism, and granitization there followed a period of relative quietness, to be interrupted by the in-

trusion of irregular dykes and sheets of a younger granite. These intrusions are particularly common in the western part of the area. They traverse all pre-existing rocks and sometimes cut from them angular inclusions. The contacts are sharp and thermal effects are negligible.

The New Granite is a fairly homogeneous, unfoliated aplitic biotite-granite. Modal analyses of specimens from intrusions all over the area are very similar, the average composition being: Quartz 30·3 per cent, oligoclase 30·5 per cent, potash-feldspar 27·8 per cent, myrmekite 5·0 per cent, biotite and chlorite 5·7 per cent; small quantities of magnetite, zircon, apatite, allanite, and secondary muscovite are present. The oligoclase forms well-defined, euhedral crystals, unlike the clumsily constructed and irregularly zoned plagioclase of the Aswan Granite.

In addition to the fine-grained intrusions, there appears to be a rare coarse-grained variety of about the same mineral composition. Only four such intrusions were found; the longest starts just north of the British military cemetery a kilometre and a half south-west of Aswan Town, and runs northward for about three hundred yards. These coarse intrusions may represent a slightly earlier phase of the New Granite, but no definite indication of their age was found.

Pegmatites and Aplites

The pre- and syn-tectonic acid veins, most of which are pegmatitic, have already been described. Pegmatite and aplite intrusions contemporaneous with or later than the granitization are also common and can occasionally be seen cutting the earlier foliated white oligoclase-rich pegmatites. Veins formed during the granitization traverse inclusions in the Aswan Granite, while those formed after granitization also cut the granite itself. The only acid veins cutting the New Granite are rare thin aplites.

A broad comparison of the pegmatite veins of various ages indicates that the younger varieties are distinctly finer-grained and more potassic than the older. The decrease in grain-size is probably due to a diminution in the gaseous and fugitive constituents which must have been largely consumed by the earlier intrusions and which helped them to travel far outward into the country rocks. The contrast in grain-size of the two granites may reflect the same change.

Late Belts of Shearing

Strong shear-lines, mostly along minor faults, are fairly widespread and have served as foci of late hydrothermal activity. In the area

south of the dam, and especially about the latitude of El Hesa Island, these late shears become more frequent and develop into wider belts of shearing. Where shear-lines pass through augen gneisses or granitic gneisses the feldspar porphyroblasts pass into porphyroclasts and the augen structure becomes more pronounced. It is often difficult to distinguish sheared migmatites from the fine grained granites interbedded with them. Near the eastern boundary of the area, basic rocks traversed by shear-lines are converted into antigorite or tremolite rocks, or into vermiculite skarns.

Post-Granitic Minor Intrusions

Small bodies of intermediate or basic igneous rock intrude all the rock types previously described. They are divisible into several kinds of lamprophyres (mostly camptonites with a red-brown hornblende), porphyrites, bostonites, diorites and dolerites, and were probably intruded at widely different dates. Some varieties, such as the diorites, appear to be older than the Nubian Sandstone, but other intrusions occupy faults (e.g., the very conspicuous fault-line north-east of, and parallel to, the dam) which cut both the migmatites and the Nubian Sandstone. There may thus be two broad sets of intrusions, the older of which perhaps represents the last manifestation of the phase of plutonic activity described above.

Although some intrusions form necks or irregular branching bodies, the majority belong to two sets of dykes, one running east-north-east and the other running north and south. The dykes usually occupy tensional joints, faults, local fractures, and other planes of weakness.

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Geology and Geography of the London–North Sea Uplands in Wealden Times

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Abstract

The uplands seem to have consisted of an interior highland north of the Thames made principally of Upper Palaeozoic rocks, and a marginal vale-and-scarp lowland of Jurassic strata on the south. This lowland bordered the Wealden morass. Much of the highland exceeded 1,000 feet. It was probably a high-level peneplain resurrected by end-Purbeck earth-movements. Two main rivers entered the swamps and account for most of the detritus. Their probable relations recall those of the contemporary rivers in Belgium.

Introduction

THE Wealden rocks are hardened river silts, clays and gravels. They accumulated as alluvial flats, deltas in freshwater ponds and lakes, muds in brackish backwaters, and so on, according to time and place. Probably they span all the time covered by the three Neocomian marine stages elsewhere (but *cf.* Wolburg, 1949)—that is from the Upper Tithonian (~ Purbeck) to the Aptian (Lower Greensand).

The inferred margin of the formation in S.E. England is shown in Text-fig. 1. This is roughly equivalent to the average shoreline of the basin, where the swamps and marshes gave way to higher ground. In places they were initially more restricted, but gradually spread further afield. Locally, the thickness of sediment eventually exceeded 2,000 feet.

The uplands that bordered the Wealden drifts were made of solid rocks and dissected by rivers carrying away their debris to the marshes. These vanished lands had, of course, their own geographies, geologies, soils, climates, plant and animal populations, and so on. Is it possible to learn a little more about these from their denuded remnants and the debris? Can we in imagination put back the Wealden detritus on to the flattened stumps and re-form the hills and valleys?

It seems we can, simply because the Wealden rocks are what they are. Accumulated under almost tideless conditions with little fetch, this sort of detritus tends to "stay put", far more so than marine sediment. In such an environment deposition is largely controlled by currents coming from the inflowing rivers. Hence the sediment laid down retains directional properties revealing past stream patterns. These in turn provide clues to the positions of the nearby estuaries. One can thus recognize the detrital matter from different catchments, and finally reconstruct the varying character of the distant land mass.

(Marine sediments, liable to drift and mix for miles along a coastline are more difficult to interpret. The Lower Greensand is one example of such.)

After making due allowance for weathering and other geochemical changes, one may therefore conclude that differences in the exotic detritus of the Wealden reflect the varying character of the uplands. And since most of the Wealden sediment of the Weald came from the lost lands over what is now the Lower Thames valley and estuary, the present paper will deal with these, the " London–North Sea Uplands ".

Broad Features of the Uplands

The London–North Sea Uplands are now represented only by the " London Platform " and its easterly extension. This is a flattish shelf of ancient rocks (Strahan, 1913; Davies and Pringle, 1913), buried beneath Lower Greensand and Gault and comprising only their roots. What little is known of its geology north of the Wealden limit is summarized in Text-fig. 1. The varying reliability of the interpretation is suggested by the distribution of borings.

A broad tract of Devonian beds covers an extensive area beneath central and western London. On general grounds and some structural and geophysical evidence (Falcon and Tarrant, 1951), it is concluded that this continues eastwards to connect up with similar rocks beneath Essex (Fobbing, Canvey Island). Southwards, the Devonian disappears under a broad unconformable ribbon of Middle and Upper Jurassic. This in turn passes under the Wealden, here with approximate conformity of strike. The Jurassic ribbon sends a tongue north-westwards into central London (pre-Gault structural depression), then passes eastwards across northern Kent. At first it lies mainly on Lower Palaeozoic outcrops (Bobbing, Chilham), completely hiding the Devonian, but steps back on to Upper Palaeozoic in the area of the coalfield. To the north, along the shore of the Thames estuary, Lower Palaeozoic rocks outcrop beneath the Lower Greensand, separating the Jurassic from the Devonian in Essex. Beyond Sheppey, Devonian and Upper Carboniferous also underlie the marine Lower Cretaceous cover, separating the Jurassic and Lower Palaeozoic outcrops. Finally, over the Kentish coalfield, a sharp difference of strike causes the whole Jurassic and some Palaeozoic to pass under the Wealden.

In Wealden times the geology of the Uplands must have been similar, but *plus the rocks removed in the Aptian interval.* Obviously, where Gault lies on the platform we have to replace more than where Lower Greensand lies on it. Now in every borehole known to the writer but two (Cliffe, Sheerness (? Fobbing)), the Lower Greensand is present only where there is Jurassic underneath. The Palaeozoic is

normally covered by Gault. Broadly, therefore, the Wealden geology was: " Palaeozoic + + ", " Jurassic + ".

This raises an interesting point about the topography of the Uplands. The Aptian Sea advanced only across the Jurassic outcrop. It was held up by the Palaeozoics. Hence the Jurassic rocks, including several very soft formations, probably formed a low vale-and-scarp fringe to the Wealden marshes. This was later easily inundated by the sea. The Devonian and Carboniferous rocks, on the other hand, are relatively hard and formed an interior highland against which the Aptian Sea battered in vain. If the rate of erosion were about the same as in Britain to-day, and the Lower Greensand interval around a tenth of Cretaceous time, the interior highland must initially have risen more than 1,000 feet above the Wealden morass, even neglecting earth-movements.

The two borings showing Lower Greensand in contact with Palaeozoic support this interpretation. At Cliffe and Sheerness the ancient rocks are Lower Palaeozoic mudstones and shales. These are relatively soft, and must have behaved like the Jurassic, forming part of the marginal lowland. That is why the Aptian Sea made its greatest inundation here. (Fig. 9 in Dewey, Pringle and Chatwin, 1925.)

Down to the south the Wealden debris from the Uplands consists of the sandy Hastings Beds below and the more argillaceous Weald Clay above. This sudden incursion of sand into southern England at the start of Cretaceous times was a momentous event, the first of of its kind for many a long day. It must indicate uplift of the long denuded London peneplain and rejuvenation of its drainage. So we get another glimpse of the Uplands. They were probably flat-topped hills intersected by down-cutting streams with terraces, nick-points and all the rest. Perhaps they resembled certain parts of South Wales to-day.

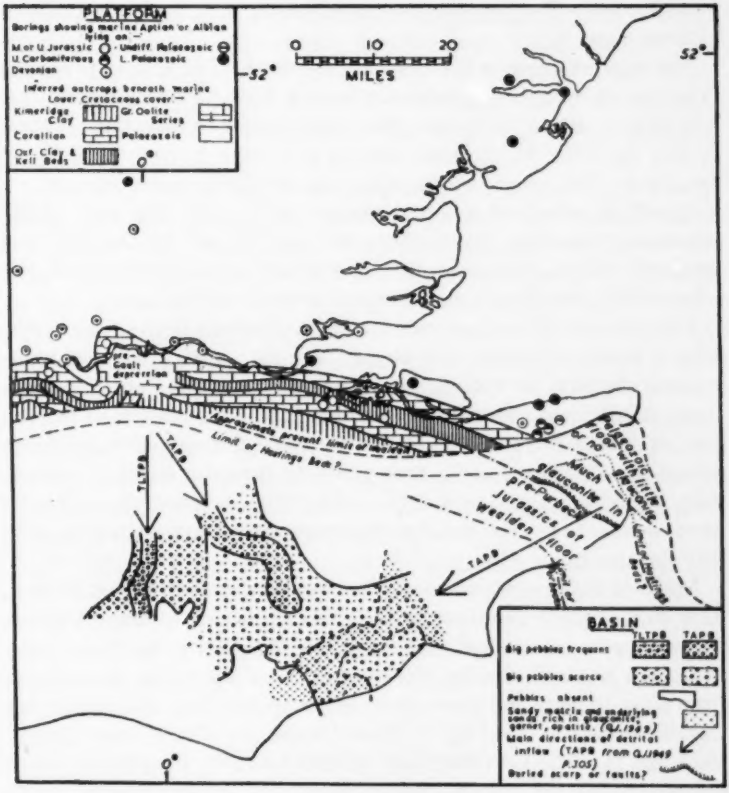
The Eastern Uplands

At first, the Wealden sediments cannot have extended far beyond the present limit of the Purbeck (Text-fig. 1). But they gradually spread further afield, and by the end of Hastings times (if lithological correlation be correct) had nearly crossed the Kentish Coalfield. Later, in Weald Clay times, they probably extended at least nearly to Thanet.

An attempt to show these stages has been made on the map (Text-fig. 1). The boundaries are reliable only in east Kent, where borings are numerous. The limits of the Hastings Beds and Weald Clay shown passing down-Channel have been inferred from isopachytes over the coalfield and the boundary in the Boulonnais.

There can be no doubt that a major river entered the Weald across this district. Immediately to the south-west, around Hastings, the rocks

show clear signs of strong currents from the north-east. Over a long strip of country inland of the town, ripple-marks, false-bedding, and so on show this very well. Often the waters were violent, for



TEXT-FIG. 1.—South-eastern England to-day. *Platform:* "Outcrops" inferred beneath Aptian-Albian cover (east Kent after Geological Survey). Distribution of borings indicates varying reliability of interpretation. *Basin*: Inferred margin at three stages, and critical petrological facies in Hastings Beds. (Sub-Wealden floor in East Kent after Geological Survey.)

bone-beds are uncommonly frequent there (1949*b*, fig. 45, p. 276). And once, when the shallow water had been colonized by horsetails, the currents scoured away all but the deepest rootlets along a N.E.–S.W. strip (1947, fig. 55, p. 305).

In this tract the sandy top Ashdown detritus is quite different from that elsewhere. It is extraordinarily rich in glauconite, garnet,

apatite and certain other minerals. The glauconite becomes less abundant in the higher beds about Hastings. Some years ago the suggestion was made (1949*a*, p. 299) that the garnet-apatite association indicated an Upper Carboniferous source and the glauconite a Jurassic one. The Jurassic source, whatever it was, clearly declined as time went on.

The tract in question is shown in Text-fig. 1. The arrow, giving the direction of inflow of the detritus during Top Ashdown Pebble Bed ("TAPB") times, is taken from the earlier publication (1949*a*, p. 305, fig. 23). As no large pebbles ever seem to have come along this route, the mouth of the river was probably some way off.

It will be observed that the stream crosses just that part of the subsequent Wealden floor where the outcrop of the Jurassic was probably widest. Moreover, the pre-Purbeck Upper Jurassic rocks are often notable for their high glauconite content in this region.

The presence of much glauconite in the Hastings Beds to the south-west is therefore readily understood. The map also explains why this mineral declines in the higher strata. Immediately after Purbeck times the Jurassic strata beyond the Purbeck limit must have formed part of the catchment. They also extended farther north-eastwards across the Palaeozoic rocks than now. In time the Hastings detritus built up across them, and their outer fringe receded through sub-aerial denudation. The outcrop thus contracted, contributing less and less glauconite.

Much of the Jurassic was soft, the overlap and overstep are proved. This part of Kent was therefore a low-lying district in early Wealden times. It may, indeed, still have something of the form of a buried vale-and-scarp lowland, for the thickness of Wealden varies capriciously from place to place and there are some unexplained bends in the harder Corallian outcrop (Text-fig. 1; from Lamplugh, Kitchen and Pringle, 1923, fig. 2). The Corallian floor is also probably irregular in detail (*op. cit.*, p. 76). This would connect it with the Bas-Boulonnais, where the Jurassic was uplifted between the Purbeck and Wealden (Pruvost, 1925).

What lay beyond the lowland? The detrital minerals suggest Upper Palaeozoic, and so does the degraded platform to-day. On the north of the coalfield both divisions of the Wealden come to lie on Palaeozoic rock. Hence, though the ancient formations here must have been blanketed by Jurassic in early Wealden times, they came to the surface not far to the north-east. Borings on the north of the Wealden boundary are confirmatory. In the group of seven on the map, Gault lies successively on Coal Measures, Devonian and Lower Palaeozoic towards the coast. "Putting back" the rocks removed during the interval, one can see, quite independently of the minerals, that this

was essentially an Upper Palaeozoic tract of country. Composed of hard rocks, it must have stood above the level of the Jurassic lowlands.

What was the actual course of the river draining this terrain? In the lower reaches it was probably more or less that of the arrow. Farther up, inland of Dover, it crosses the sinuous Corallian outcrop already mentioned. Lamplugh, Kitchen and Pringle tentatively explained these irregularities as due either to a buried scarp or to (? pre-Wealden) faults (1923, p. 15). In either case they may be related to the Wealden river. Perhaps one of the bends might even mark the actual valley.

Above the Corallian the river traversed the Dover district. Clean gravels and sharp sands are unusually frequent in the attenuated Wealden there, and the pebbles reach sizes unapproached anywhere in the Weald. Nearly all the pebbles are of white quartz, but reddish quartzite or quartz is known (Lamplugh, *in* Lamplugh and Kitchen, 1911, p. 19). Chert and glauconite are virtually absent. All this is consistent with derivation from Upper Carboniferous and Devonian outcrops upstream of the glauconitic Jurassics.

We are now well inside the narrowing " embayment ". Further on, one can only surmise. But clearly, if the reconstruction of the embayment is substantially correct, the valley almost certainly took a sharp bend and headed into the Palaeozoic hills via Thanet. Hence, perhaps, the gravels at Maydensole and Ebbsfleet.

As far as can be seen, the relations of the Kentish Wealden strongly recall, on a larger scale, the situation proved in Belgium further south-east (Stevens and Marlière, 1944; Marlière, 1934, 1946). Between Bernissart and La Louvière at least three deltas are known, and they all lie in the V-shaped open ends of south-sloping valleys carved out of Upper Carboniferous rocks.

The Central Uplands

Between Kent and London no major inflow yields a clue to the constitution of the northern hills. Only the borings into the platform and some indirect evidence are available. This suggests, as already seen, that the Jurassic and Lower Palaeozoic strata formed a marginal lowland to the swamps, extending as far as the line of the Thames estuary. Beyond, rose the Palaeozoic uplands of Essex, destined not to be submerged by the sea until Gault times. As the ancient rocks in the Fobbing and Canvey Island bores are Middle Devonian or later, these hills must have included members of younger Systems. Probably remnants of a Jurassic cover persisted here and there.

More than this cannot be said. The lateral extents of the Middle and Upper Jurassic transgressions are unknown, and detritus from the Essex hills has yet to be recognized in the Wealden.

The Western Uplands

In many places the Hastings Beds show signs of a second river entering from the north-west. Ripple-marks, false-bedding, much contemporaneous erosion in the north-west (1949*a*, p. 271) and so on, amply prove this.

At the close of Ashdown times a thin sheet of gravel spread south-eastwards across the basin. The boundaries of this are shown on the map (Text-fig. 1). Within it, the pebbles get bigger and more abundant north-westwards. At the same time their matrix becomes finer and more deficient in glauconite. This was obviously because of increasing distance from the mouth of the sandy Kentish river. The biggest pebbles are commonest in a narrow meandering belt, as shown. Taken with the outer boundary of the bed, this establishes the presence of a river inflow to the north-west, in the general direction of central London. An arrow (labelled " TAPB "), taken from the earlier publication (1949*a*, p. 305, fig. 23), illustrates the point. Since pebbles were discharged into the area, the mouth of this river was presumably nearer the Weald than that of the Kentish river.

During recent years other horizons of the Hastings Beds have been mapped in a similar way. The most interesting has proved to be a second pebble bed, about 200 feet higher, capping the Lower Tunbridge Wells Sandstone. This is almost identical with the earlier one, both in lithological relations, types and proportions of constituents, and so on. Indeed, the two are so alike that *there can be no question of their origin in the same catchment.*

The distribution of the Top Lower Tunbridge Wells Pebble Bed is superimposed upon the same map as the other (Text-fig. 1). It outcrops to the west of the older bed and its pebbles clearly rolled down from the north (arrow, marked " TLTPB "). This gives a clue to the position of the estuary, for the river must have entered somewhere near the intersection of the two inflow-directions. The exact position depends on the form of the gravel spreads. If the bigger pebbles actually floor underwater channels continuous with the principal deltaic pass of the time, then the apex of the delta probably lay close to the intersection. This apex would be somewhere near West Wickham, well south of the Wealden feather-edge. The position of the main pass below West Wickham would thus have changed during the Ashdown–Tunbridge Wells interval.

If, on the other hand, the coarser gravels floor minor distributaries within the delta, the apex more likely lay rather north of West Wickham.

Almost certainly, the first interpretation is correct, for the gravels seem to fill broad shallow channels. Thus the basal Wadhurst Clay and the Grinstead Clay, overlying each respectively, are both thickest where

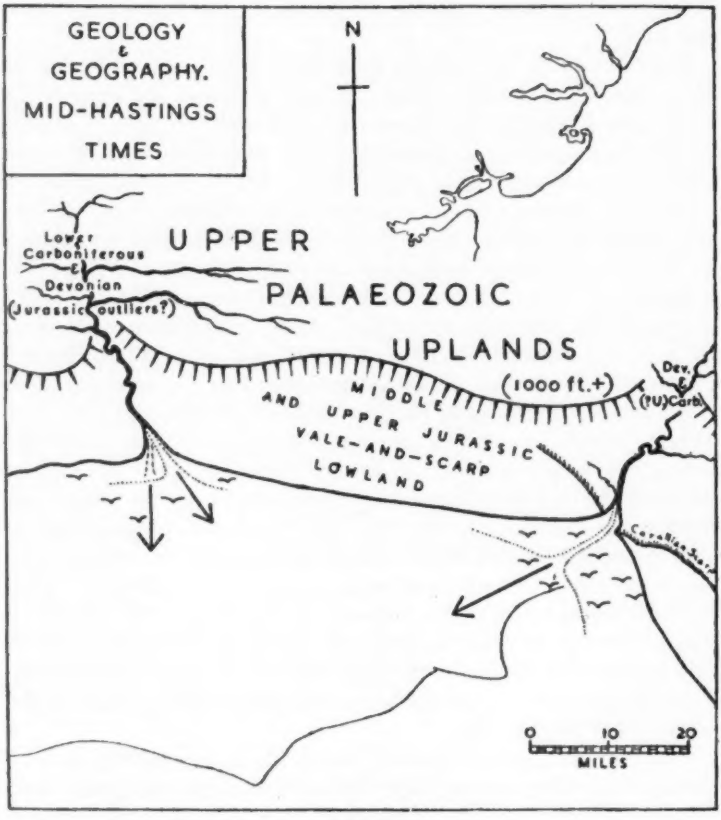
the gravel is coarsest (1949*a*), pp. 301–3; *cf.* Text-fig. 1 with Fig. 30 in Reeves, 1949, p. 257); and the coarsest spreads are 2–3 miles wide.

Why then should the river mouth have lain south of the present limit of the Wealden, which is known to be somewhere between Addington and Streatham? The answer is that we are dealing with Hastings Beds, and here these must be overlapped by the higher beds, as in east Kent.

Investigation of the pebbles themselves reveals something of the rocks exposed upstream. Details of their petrology are reserved for a later paper. Only broad features will be mentioned here. The commoner types comprise sedimentary cherts, red-stained and unstained rocks of ancient aspect, and phosphorites. Very rare bits of oolite are noteworthy among the less common sorts (Milner, 1923, p. 288). The phosphatic pebbles undoubtedly came from Jurassic formations, for some are casts of ammonite chambers (Topley, 1875, p. 84; Kirkaldy, 1947, p. 235), and others enclose foraminifera. Possibly also the oolitic fragments are Jurassic. Many of the sedimentary cherts can be closely matched with Lower Carboniferous types in the Midlands and North, and one has yielded a diagnostic fossil. A high proportion seems to be secondary after partially dolomitized limestone. The red-stained pebbles are chiefly crushed and quartz-veined quartzites, jaspers, acid lavas and so on. A few are slightly rounded split pebbles, obviously derived from pre-existing conglomerates. In detail, the pebbles recall varieties in the Old Red Sandstone conglomerates of western England (Wallis, 1928). No tourmalinized rocks are present, so they are unlikely to have come from Permo-Triassic sources. Hence they were derived from the Devonian on the platform. The importance of these outcrops increased as time went on, for the red-stained pebbles are a little more abundant in the Tunbridge Wells rocks. Significantly, in the Ashdown pebble bed they are sensibly more frequent towards the north-west (1949*a*, p. 268).

A look at the London Platform shows the reason for this suite of pebbles. Any river coming from the north-west or north must have crossed a broad strip of Jurassic rocks. These extend to-day as far as the line of the Thames. The phosphatic pebbles (Oxfordian?) and fragments of oolite (Great Oolite?) are thus accounted for. Beyond the present Thames the Devonian core, including Old Red Sandstone conglomerates, stretches for 15 miles or so, and may have reached Ware (Wenlock under Gault) in Wealden times. The strata are late Devonian (or early Carboniferous?) beneath Tottenham Court Road, Willesden, Southall and Cheshunt (Turnford). In the last three places they are overlain by Gault, and at Tottenham Court Road by Great Oolite. There is therefore every likelihood that thin Lower Carboniferous outcropped in the Western Uplands, even if the Jurassic

had at times crossed the platform. On general grounds, we should expect developments of silicified and dolomitic limestone. Though no Carboniferous is known in the London district, it is worth noting that at Cambridge it is almost entirely dolomitic (Pugh, 1954, p. 11),



TEXT-FIG. 2.—South-eastern England in mid-Hastings times.

while certain of the Upper Devonian limestones at Willesden are also dolomitized (Stubblefield, 1948, p. iii).

I believe (in extension of Milner's views, 1923, p. 298) that the pebbles in the Wealden clinch this matter, and that we may visualize the London Uplands as showing inliers of Devonian in the valley bottoms with Avonian and Jurassic on the higher ground. As the Carboniferous and Jurassic were denuded away, more Devonian was

exposed, accounting for the slight upward increase of red-stained quartzites observed in the Hastings Beds.

What course did the river take? The stream-patterns in the basin perhaps suggest that its valley ran north-north-westwards from the mouth. Once again, as in Belgium and possibly in Kent, the route traverses a structure representing some sort of " embayment ". This is the pre-Gault downwarp or fault-depression proved by the surprising tongue of Jurassics at Tottenham Court Road. Its age and exact nature are, of course, uncertain, but there can be no denying the coincidence. And in view of the Wealden valley structures known elsewhere, it must be suspect as marking the structurally guided course of our stream. The oblique entry of the river into the basin would then be understood.

With removal of the Jurassic cover, the upper part of the catchment will have been an area of superimposed drainage.

Submergence of the Uplands

Eventually inundation by the Lower Greensand Sea destroyed the Wealden environment. At first the flooding did not extend much beyond the lowlands. Thereafter the sedimentary messages from the nearby lands are confused and difficult to decipher. Like most seas, the Aptian Sea mixed up materials from far and near. Finally, in Gault times, the whole region was submerged; and the messages cease.

Summary and Prospect

A rough reconstruction of the Wealden geography and geology of the London–North Sea Uplands is given in Text-fig. 2.

Other kinds of map are at present out of the question. We cannot clothe the hills in vegetation or people their environments with animal life. Nor can we say much about the soil of the Uplands, or of the weather overhead. Clearly the climate was not monsoonal, for lateritic materials do not occur. All in all, the molluscs indicate weather not very different from that in the Weald to-day. Yet weathering seems to have been intense, for virtually only siliceous detritus survives. Perhaps we must invoke long-continued leaching on the end-Jurassic peneplain as the cause.

These and related questions are unanswerable now. But they point the way to future endeavour.

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UNIVERSITY OF READING.

The Structure and Development of *Peiragraptus fallax*, gen. et sp. nov. A New Graptolite from the Ordovician of Canada

By Isles Strachan

Abstract

A new genus, *Peiragraptus*, is proposed for graptolites having a partial diplograptid development and mainly uniserial rhabdosome. A growth series of the new species *P. fallax* from the Upper Ordovician of Anticosti is described.

AMONGST the specimens from Canada in the Lapworth Collection is a piece of limestone from Observation Cliff, Anticosti, on the surface of which numerous fragmentary graptolites are to be seen. They are mainly uniserial but close examination reveals the presence of a small theca at the proximal end suggesting the beginning of a second thecal series. In spite of the rather poor preservation, a small chip of the limestone was treated with acid and a large number of specimens obtained from the resulting muddy residue. These have proved to be a new form of graptolite and although its exact horizon is unknown, it seems of sufficient interest to warrant description. All the specimens from the limestone appear to be of the same form and a more or less complete series of growth stages has been obtained. If only one or two specimens had been found they might well have been regarded as simply malformed diplograptids but since some fifty specimens have been isolated and there are as many again still visible on the original block, this peculiar graptolite, even if only aberrant, was certainly capable of reproducing itself and can fairly claim specific status. There is no place for it in the present genera of graptolites and so a new genus is here proposed for its reception.

Family Diplograptidae

Genus *Peiragraptus* gen. nov.

Genotype.—*P. fallax* sp. nov.

Diagnosis.—Rhabdosome uniserial except for the presence of th1[2], development of incomplete diplograptid type.

The other characters of the genus are at present those of the unique species. The essential character is the failure to develop the second thecal series after the initial diplograptid growth stages. If further species are found, it may be necessary to introduce the form of the theca into the definition of the genus as is done in other diplograptid genera.

Peiragraptus fallax sp. nov.

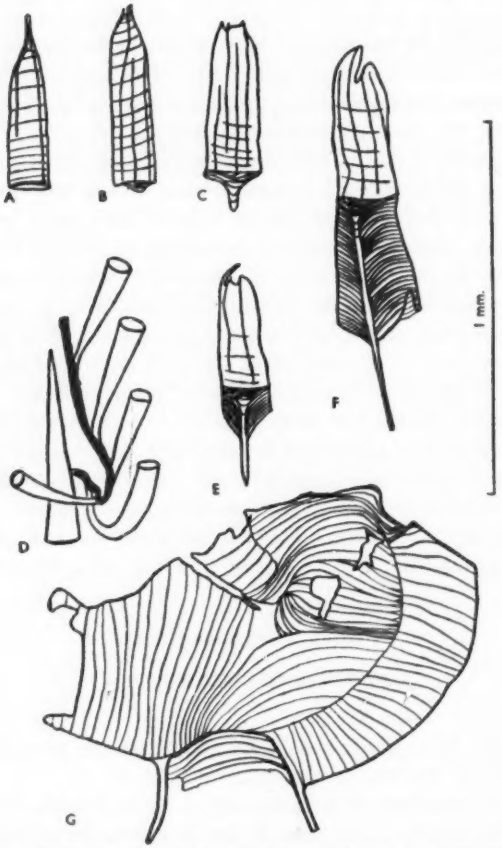
Holotype.—B.U. 682 (Text-fig. 2*H*).

Horizon and Locality.—Observation Cliff, Anticosti, probably from Vaureal Formation, Upper Ordovician (Twenhofel, 1928).

Derivatio nominis.—πειρα = an attempt.
fallax = deceitful.

Rhabdosome uniserial except for the proximal thecal pair; up to 1 cm. long and 1 mm. broad; thecae 16/cm., with the apertures opening into slight excavations.

The prosicula is a simple cone, 0·4–0·5 mm. long and 0·1 mm.



TEXT-FIG. 1.—*A, B, C*: Prosiculae and beginning of virgella (B.U. 671–3).
D: Thecal diagram.
E, F: Young stages showing growth of metasicula (B.U. 674–5).
G: Broken proximal end showing the growth bands of th1[2] and 2[1] (B.U. 676).
All × 50.

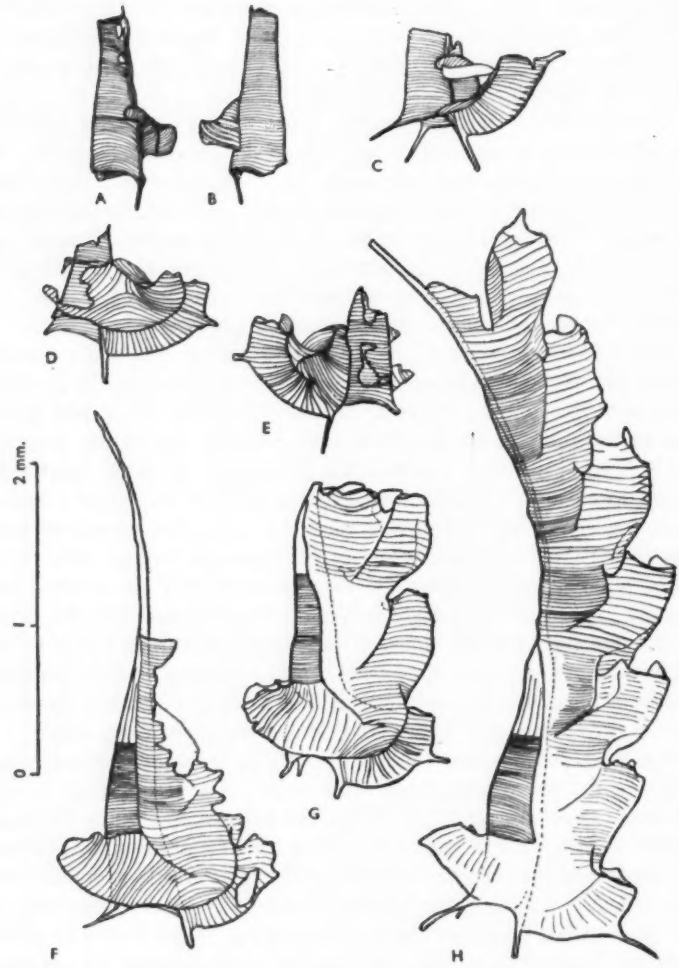
broad, usually with a well-marked spiral line, either dextral or sinistral, and, when fully developed, with two sets of longitudinal fibres, one set of which is apparently prolonged into the virgula. In most specimens the apical portion is broken (Text-fig. 1 *A–C*).

The metasicula develops in the usual manner but the first few growth bands are confined to one side of the prosicula (Text-fig. 1*B*) and form a virgella from the start (cf. *Climacograptus inuiti* Cox). The zig-zag suture is clearly seen in the early part of the virgella but secondary thickening obscures the trace of the growth lines on the later part (Text-fig. 1 *E*, *F*). When fully developed, the metasicula is just over 1 mm. long and has an apertural breadth of 0·3 mm., excluding the pair of well developed apertural spines.

The initial bud develops by resorption before the sicula is fully grown (shown by the absence of the apertural spines) and the development is at first that of a typical diplograptid. The initial bud grows first across the virgella and then down towards the sicular aperture but very soon develops a wide hood (Text-fig. 2 *A*) which marks the beginning of th1^{2}. The developmental series from this point is rather poor, but shows that the growth of th1^{2} is very slow compared with that of th1^{1}. The growth bands of th1^{1} become distinct from those of the hood rather low down on the sicula and then its outward and upward growth is rapid while it retains a connection with the hood on its inner side (Text-fig. 2 *E*). A mesial spine is formed a little below the aperture of th1^{1} and apertural lappets are also found in the fully developed theca in most specimens. The number of growth bands in the free portion of th1^{1} varies from about twenty to twenty-six.

The slow development of th1^{2} appears to be connected with the origin of th2^{1} from almost the same point. The hood grows across the initial bud back on to the sicula, but appears to stop at the stage shown in Text-fig. 2 *C* where the hood forms a connection between the sicula and the inner wall of th1^{1}. The later development of both th1^{2} and th2^{1} is from a flange which develops along the inner edge of th1^{1} and reaches across to the sicular aperture. A few (seven or eight) complete bands form a wall with a curved edge dividing the aperture below the hood into two parts and the growth bands in this central area of the reverse side fall into three distinct groups, the upward growing flange and the two laterally growing thecae, the initial part of the th2^{1} being rather narrower than that of th1^{2} (Text-figs. 1 *G*, 2 *D*).

Theca 1^{2} continues growing across the sicula but it becomes distinctly narrower in its free portion and although it develops apertural lappets it has a stunted appearance compared with the other thecae. In some specimens the aperture of th1^{2} is very contracted and this gives the rhabdosome a completely uniserial appearance, particularly in those specimens partially embedded in the limestone.



TEXT-FIG. 2.—*A*, *B*: Reverse and obverse views of a metasicula showing the initial bud and the developing hood (B.U. 677).

C: Growth stage showing th1^{1}, the complete hood and the beginning of the flange on the reverse side (B.U. 678).

D, *E*: Reverse and obverse views of later growth stage showing th1^{2} and the beginning of th2^{1} (B.U. 679).

F: Reverse view of a broken specimen showing the dorsal part of th3^{1} (B.U. 680).

G: Similar but more complete specimen (B.U. 681).

H: Largest isolated rhabdosome, to th7^{1} (B.U. 682). Holotype. All × 20.

The growth bands of the initial part of th2¹ are very variable and indicate rather uneven growth in some specimens (Text-fig. 1 *G*). Unfortunately most of the larger specimens have not bleached at all well in this region and the apparent simplicity of the growth bands in these (e.g. Text-fig. 2 *G*) may be illusory. Certainly, between the hood and the outer flange, there is very little room for the initial part of th2¹ so that it is forced to grow laterally and outward until it has passed the hood and it is then free to fill the space between th1¹ and the sicula.

The development of the later thecae from th2¹ appears to be similar to that in monograptids and septate diplograptids, each theca developing from the dorsal part of the preceding theca. The growth lines are crowded in the proximal part of each theca but become well spaced in the distal part. As is usual, there is a reduction in the number of growth bands in the distal part of the later thecae although the thecal length remains the same. It is noteworthy that except occasionally in the development of th3¹ from th2¹ there appear to be no intercalated growth bands on the dorsal side of a theca where the succeeding theca will develop (Text-fig. 2 *F*). The lower edge of the interthecal septum is clearly marked on the outer wall and the growth lines of the new theca are quite distinct from those of its " parent " from the beginning. There is thus no evidence of a " common canal " although something of that nature was probably present in the soft parts.

In the structure of the proximal end as far as th2¹, *P. fallax* is quite typically diplograptid. The adult thecae are similar to those of some orthograptids and this species can be simply regarded as a fully septate diplograptid in which th2², and consequently the second thecal series, has failed to develop. It would also be possible to derive *Peiragraptus* from a dicranograptid by similar means but it may be noted here that in those species of *Dicellograptus* which have the sicula partially embedded in one of the stipes, the sicula appears to be always attached to the second stipe, th2², etc., and not as here to the first thecal series.

The stunted appearance of th1² in most specimens suggests the presence of some inhibiting factor on that side but it seems unlikely that this was due to the mode of life of the colony as there is no sign of any features unusual in graptolites.

Although occurring at a suitable horizon, *Peiragraptus* cannot be regarded in any way as ancestral to the monograptids since, even with the elimination of th1², the development of th1¹ would require drastic modification to reach the monograptid type.

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The Oligocene-Miocene Boundary on both sides of the Atlantic

By C. W. Drooger

Abstract

The sequence of *Miogypsina* species in the type region of the Aquitanian-Burdigalian offers a good basis for correlation. Early Aquitanian corresponds to the range of *M. gunteri*, early Burdigalian to those of *M. irregularis* and *M. burdigalensis*, early Vindobonian to that of *M. cushmani*. In the western Mediterranean area application of this sequence is considered justifiable. Comparison with the similar sequence in America shows a much older occurrence of equivalent species in the western hemisphere than in Europe. A correction of the American time scale is considered premature when made on the Miogypsinidae alone.

FOLLOWING the recent discussion of the interpretation of the Aquitanian and connected problems by Drs. Eames and Stainforth, I would like to add some comments and new data.

The ascending succession of Rupelian-Chattian-Aquitanian-Burdigalian-Vindobonian is commonly considered to cover the time range from Middle Oligocene to Middle Miocene. It is generally assumed that the individual time ranges fit nicely together, without leaving gaps or showing overlap. On closer examination, however, this appears to be an over-simplified view ; but discrepancies are difficult or impossible to tackle with our present means. Type localities of successive stages are mostly irregularly scattered on the map of Europe, so that the law of superposition cannot be applied. We must therefore look for index fossils, but in this we are severely hampered by facies differences of the various type sediments. Moreover, migration times of such organisms, different survival periods, and somewhat different evolution rates in separate regions, tend to complicate matters still further. The exact relations between the components of our time scale may remain obscure, especially because of the relatively small amount of errors. Consequently, we must accept the idealistic picture of the time scale. A possible time-stratigraphic gap between two successive type sections, once proved, must be considered to belong to one of the stages, if it is not wide enough to create a new name for it. On the other hand, in the case of overlapping, one of the stages must be corrected in order to get rid of the part twice accounted for.

On this theoretical basis, I consider the time directly preceding the fairly well-defined Aquitanian as Chattian, though we do not know as yet the level that corresponds to the upper limit of the type Chattian in the Aquitaine basin. The same reasoning is applied to the Burdigalian-Vindobonian boundary.

As regards the Oligo-Miocene stages, Aquitanian and Burdigalian

offer a good starting point, since in them we have at least two stages that have been defined with clear interrelation in a single area. For their interpretation in other regions I am inclined to give special weight to certain groups of Foraminifera, though this opinion may not be free from the bias of a micropaleontologist. The " orbitoidal " larger Foraminifera and the pelagic smaller ones are important not merely because of their usually world-wide distribution, but especially their distinct evolutionary trends and their capacity for rapid dispersal respectively.

As to the Oligocene-Miocene type sections, only those of the Aquitanian and Burdigalian contain " orbitoidal " larger Foraminifera. Unfortunately their deposits of fore-reef facies contain hardly any planktonic species. Mr. Kaasschieter only found some scattered individuals of *Globigerina bulloides* and *Globigerinoides trilocularis* in the sediments of both stages.

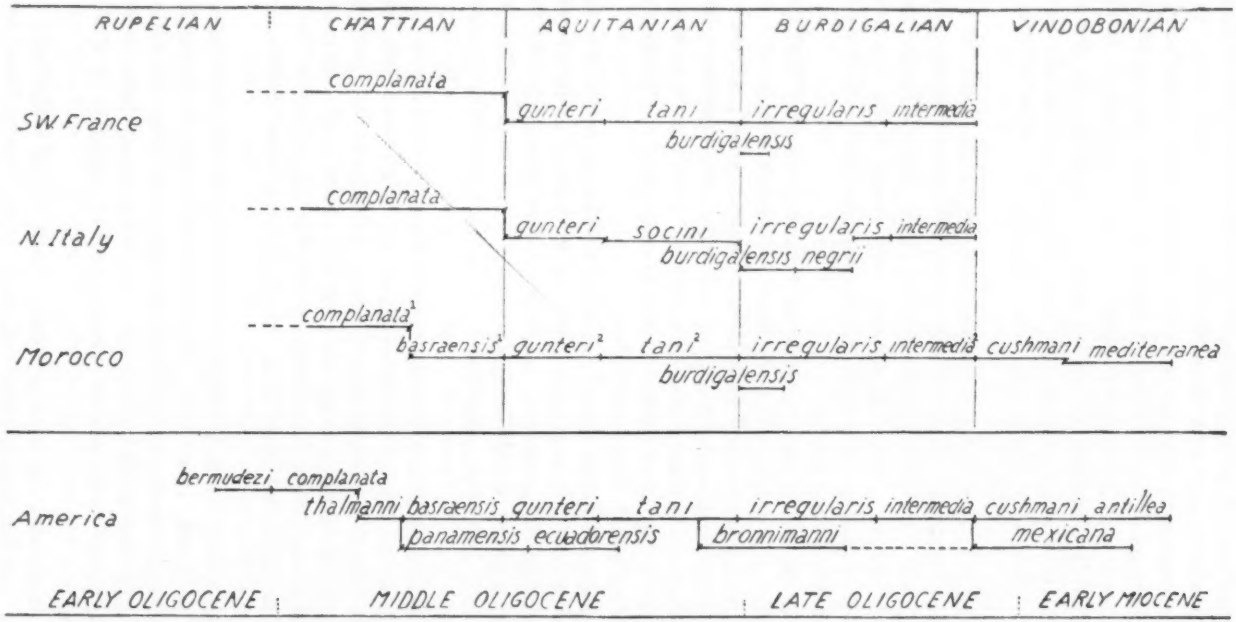
The type deposits of the Aquitanian near Bordeaux overlie Oligocene beds that contain no larger Foraminifera. At the type locality (Moulin de Bernachon, near Saucats) the lowermost Aquitanian of lacustrine to brackish-water facies is also devoid of these organisms, but *Miogypsina* (*Miogypsina*) *gunteri* was found in equivalent beds some kilometres away (railway cutting of Labrède). The younger part of this stage is characterized by *M.* (*Miogypsina*) *tani.*

The more primitive *M.* (*Miogypsinoides*) *complanata*, met with in the southern part of the Aquitaine basin, is considered to be of pre-Aquitanian age (Chattian and possibly younger Rupelian). At the localities involved (Saint-Étienne-d'Orthe, Saint-Géours, Christus, Abesse) it is always accompanied by *Lepidocyclina* (*Nephrolepidina*) *tournoueri.*

The Aquitanian is separated by a short hiatus from the overlying Burdigalian. In the older Burdigalian, the *Miogypsina* sequence continues with *M.* (*Miogypsina*) *irregularis*, followed by *M.* (*Miogypsina*) *intermedia* in the younger part. In the oldest beds *Miogypsina* (*Miolepidocyclina*) *burdigalensis* also occurs, but this species evidently entered the Aquitaine basin as an immigrant, and soon disappeared again.

The beds overlying the Burdigalian in S.W. France, which are placed in the Vindobonian (Helvetian), contain no more larger Foraminifera. But more highly developed species of the *Miogypsina* s. str. series have been found further south in Morocco : *M.* (*Miogypsina*) *cushmani* and *M.* (*Miogypsina*) *mediterranea.* These species I regard as of Vindobonian age.

These data from the Aquitaine basin allow of the following conclusions. The Rupelian-Chattian boundary cannot yet be found with the help of the Miogypsinidae. *M. gunteri* begins its range with the



TEXT-FIG. 1.—Tentative correlation-chart of the Miogypsinidae in the Western Mediterranean area and America. For each geographic area the representatives of the subgenera *Miogypsinoides*, *Miogypsina*, and *Miolepidocyclina* (with *Miogypsinita*) are given on separate levels from top to bottom.
(1) Not autochthonously observed; (2) traces only.

oldest Aquitanian. The Aquitanian-Burdigalian boundary coincides with that between *M. tani* and *M. irregularis*, while *M. burdigalensis* also marks early Burdigalian. The boundary between Burdigalian and Vindobonian can best be placed between *M. intermedia* and *M. cushmani*.

For correlation within the western Mediterranean area the Miogypsinidae appear to be very useful. The sequence of *Miogypsina* species in northern Italy is similar to that in S.W. France. In Italy the *Miolepidocyclina* lineage developed from *M. gunteri*, but it was later on replaced by *Miogypsina* s. str. The latter ended here at the same morphological level as it did near Bordeaux (*M. intermedia*). Correlation of the strata of both areas is strengthened by the migration of *M. burdigalensis* from Italy in a western direction to the Aquitaine basin and also to Morocco.

A comparable sequence of Miogypsinidae occurred in Morocco. Unfortunately it is nearly entirely obscured in the faunas described by Senn and Brönnimann, for they described associations that contain up to 90 per cent of drifted elements, a fact which could not at that time be appreciated by these authors. Probably the *Miogypsina*-bearing beds are of Aquitanian to Vindobonian age, containing the autochthonous series *M. tani* to *M. mediterranea*.

From these data on the larger Foraminifera of this side of the Atlantic it may be seen that the sequence advanced in the beginning of this century by Lemoine, R. Douvillé and others was erroneous in several of its details, and this accounts for many wrong correlations in other parts of the world.

Turning our attention now to the other side of the Atlantic, we find a *Miogypsina* sequence closely resembling that of France, though less well-founded because of the lack of a sufficient number of long, *Miogypsina*-bearing, stratigraphic sections. Comparison of the sequences clearly shows that corresponding species are estimated to be much older in America than they are in Europe. So, if other data are found that corroborate this picture, the American time scale must be considered to be wrong to a considerable extent. However, such data on other Foraminifera are still very scarce this side of the Atlantic. Planktonic Foraminifera may indicate a divergence in the same direction, but in Europe, exact details concerning these fossils in the *Miogypsina* sequence are unknown to me. I am completely at a loss for instance to know where exactly *Orbulina* begins its range on our side of the Atlantic. The stratigraphic distribution of *Lepidocyclina* is also a warning against placing too much reliance on the Miogypsinidae alone, as index fossils for trans-Atlantic correlation. Both in France

and Italy the last Lepidocyclinae are found accompanying *Miogypsina complanata*, but in America they are found at least up to the level of *M. intermedia*. This may be due to a longer survival of the Lepidocyclinidae in the western hemisphere, but such is merely a hypothesis at the moment.

As a result, change of the American Oligo-Miocene time scale and so of the place of the Oligocene-Miocene boundary is looked upon as premature on the basis of the Miogypsinidae exclusively. However, they certainly suggest a considerable shift, the exact extent of which must be determined by additional detailed data on other Foraminifera.

A final remark may be made about the placing of the Aquitanian in either Oligocene or Miocene. So far I have placed it in the latest Oligocene. However, with regard to its microfauna at the type locality, there is but slight difference from that of the type Burdigalian. Though of course for the greater part based on facies resemblance, this fact favours a grouping with the Miocene. In that case the natural limit suggested by the Miogypsinidae is doubtful : some species of *Miogypsina* s. str. (*M. thalmanni*, *M. basraensis*) are likely to be pre-Aquitanian so that the change from *Miogypsinoides* to *Miogypsina* s. str. does not mark the Oligocene-Miocene boundary. *Lepidocyclina* may be no more helpful either. In my opinion the placing of the Aquitanian is not so important, so long as it is correctly interpreted.

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CORRESPONDENCE

THE USK MONCHIQUITE—A " PIPE INTRUSION "

SIR,—The discovery of the Usk monchiquite is reputed to be due to a former Cardiff student who noted a very black rock being used as a roadstone on a private drive. He took specimens to Professor W. S. Boulton, then Professor at Cardiff, who gave an account of the occurrence (*Quart. Journ. Geol. Soc.*, 1911, lxvii, 460).

The present author made a number of visits with student parties during the period 1920–1938. There was then only a small exposure at the side of a private drive, with the rocks almost concealed by vegetation.

When class excursions were resumed after the second war, it was found that a quarry, about 200 ft. long, had been opened on the monchiquite outcrop, presumably to obtain material for camp roadways, other hard stone being scarce in that part of the Old Red Sandstone outcrop.

The higher side of the quarry wall then afforded a fine vertical section of the monchiquite intrusion, but the floor of the quarry was concealed by a great mass of large blocks of unweathered monchiquite liberated by blasting.

The quarry face consisted partly of unweathered and partly of weathered material. The junctions of weathered and unweathered material were exceedingly irregular, and in places the original intrusive rock had weathered to a reddish-brown soil, with included blocks showing all stages from the fresh black rock to the brown soil.

The new and larger exposures sufficed to show that the occurrence was a pipe-intrusion exactly similar to occurrences of kimberlite pipes in the Kimberley area of South Africa.

The similarity to the South African pipes is, of course, no guarantee that the Monmouthshire intrusion carries any material of economic importance. There are plenty of " pipes " in South Africa which fail to yield any minerals of commercial importance.

Along a large part of the exposure in the quarry wall the intrusive rock extended right up to the surface. But on one side of the excavation the igneous rock was overlain by a deposit of well-rounded pebbles which must have come from a considerable distance.

A more recent visit, 1954, showed considerable changes in the appearance of the ground in the vicinity. The former extensive woods had all been cut and turned into agricultural ground, so that it would now be difficult to recognize the locality from Professor Boulton's original description, though the newer pre-1948 quarry is still visible on a cleared hillside.

" TAN-Y-RHIW,"
RHIWBINA HILL,
NR. CARDIFF.

A. HUBERT COX.

19th October, 1954.

THE GREAT GLEN FAULT, CO. DONEGAL

SIR,—In a recent issue of the *Magazine* (Vol. xci, p. 338), J. B. Auden poses the question of the extension of the Great Glen Fault into Co. Donegal. In the process of surveying a large tract of the western and northern parts of that County we have already accumulated considerable evidence of the presence of a major S.W. trending fault.

Such a dislocation can be traced from the northern tip of Inishowen, along the northern side of the Knockalla Mountains, through the Leannan valley and through the localities of Drumbologe Bridge, Swilly Bridge, Cummirk Bridge, and thence presumably by Lough Ea and Sir Albert's Bridge. In Inishowen it constitutes the smash zone of W. J. McCallien (1935, *Proc. Roy. Irish Acad.*, xlii, p. 407), north of Knockella it forms the faulted boundary of an outlier of presumed Old Red Sandstone age (H. E. Wilson, 1953, *Proc. Roy. Irish Acad.*, lv, Sec. B, No. 13, p. 294), near Milford two very

different Dalradian quartzites are brought into line and thence for many miles to the south-west, garnetiferous schists are brought against rocks in a very different metamorphic condition.

The rocks on either side of this fault show strong contrast in sedimentary facies, structure, and metamorphism. Where we have mapped this important line in detail it consists of a number of closely spaced but slightly curving faults which dissect the country into a series of lenticular strips. There is additional evidence of movement in the presence of both coarse and fine grained crush breccias, veining, and slickensiding. Parallel to the main dislocation there are numerous subsidiary faults, some of which show considerable strike-slip movement ; we may mention the Gweebarra and Lough Belshade Faults, the latter with a left-handed strike-slip of 2·2 miles (G. Walker, 1954, *Geol. Mag.*, xci, p. 116).

We have still considerable areas to survey and it will be some time before we can display this structure in full. Until then, we shall not hazard a guess as to the extent or sense of relative movement. We can, however, affirm that this Knockalla-Leannan Valley Falt is a major dislocation which is in direct line with the Great Glen.

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10th November, 1954.

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R. M. SHACKLETON,
W. S. PITCHER,
R. L. CHEESMAN,
H. H. READ.

REVIEWS

OIL IN THE MIDDLE EAST. By STEPHEN HENRY LONGRIGG. pp. xiii + 305. Oxford University Press (under the auspices of the Roy. Inst. Internat. Affairs), 1954. Price 25*s*.

Industrial enterprise, guided by geological science and its physical aides, has discovered vast reserves of oil in the Middle East ; increasing world demand for petroleum products will cause this process to continue. As geologists know, the formation, concentration and preservation of this enormous amount of oil is due to a combination of exceptionally favourable geological circumstances. Owing to a scarcity of suitable publications, they are less well informed on the history of the oil operations and the political stresses involved in the creation and distribution of this wealth. Brigadier Longrigg, well qualified for the task, has written (under the auspices of the Royal Institute of International Affairs) a book which can do much to remedy this. It is not a geological book, but includes a general geological introduction and brief notes on the geological results of many test wells. All geologists interested in the results of their professional activities could profitably read it ; and give some thought to the statesmanship required of many nationalities if lasting benefits are to result from this modern, oil soaked, Eden. The book has five maps showing the distribution of oilfields, pipe lines, concession boundaries, etc. ; it seems unnecessary to have merged all the main Persian fields into one monster on the first map, causing it to overflow one concession boundary in two places !

N. L. F.

ALBERT HEIM. LEBEN UND FORSCHUNG. By MARIE BROCKMANN-JEROSCH, ARNOLD, AND HELENE HEIM. Wepf and Co., Verlag, Basel. 268 pp., with 11 figs. and 15 plates. 1952. Price S. Fr. 18·50.

Albert Heim died in 1937 in his eighty-ninth year. His life spanned the great period of world-wide geological exploration and to Heim must for ever

be the credit for resolving so many of the mysteries of Alpine structure, which have had such a profound influence on geological thought during his life-time and since. His greatest works are *Mechanismus der Gebirgsbildung*, published in 1878, and his monumental *Geologie der Schweiz*, which appeared between 1919 and 1922.

This biography of Albert Heim is a worthy memorial to a great scientist whose interests and influence ranged far beyond the bounds of his principal subject. He was unusually talented, in exposition, composition, and illustration; his long tenure of the Chair of Geology at Zürich, from 1874 to 1911, was truly an epoch. His family life, his personal development and evolution, and his manifold outside activities are all fully and sympathetically told and are an important record, but to geologists the world over the important chapter, and this all too short, is that in which is described the part which Heim played in the development of ideas on the fundamental tectonics of the Alps.

The existence of overthrusts in the Alps was first recognized by Arnold Escher von der Linth in 1841 by the superposition of Cretaceous over Tertiary in the Canton Glarus and, south of Wallensee, of the succession, Cretaceous to Trias, over nummulitic beds. To the end of the century the Glarner Alps became the focal point of tectonic controversy. Escher thought that, contrary to the normal pattern of the Alps, the Glarner thrust was towards the south. Heim resolved the difficulty by assuming two folds, recumbent and thrust towards one another, almost meeting and enclosing the Tertiary syncline. In 1884 Marcel Bertrand suggested that the double fold was in fact one simple thrust sheet, but, as he had not even visited the area, Swiss hackles were raised and this explanation was discountenanced. In 1902 Maurice Lugeon produced his brillant work, *Les grandes nappes de recouvrement*, and described so many examples of thrust sheets along the length of the French and Swiss Alps that the famous Glarus example took its place modestly as one among many. The dramatic moment came in the following year, 1903, at the International Congress in Vienna when Lugeon lectured on his theme. Turning towards his old master, Albert Heim, he said:—

" Theories are ephemeral, facts are permanent. And if these had not been assembled to make a firm background I would not have been able to interpret your magnificent mountains otherwise than you. Among all who have written on the famous Glarner double-fold, you were nearest to the truth. You are our master. You have shown us the new way to modern tectonics."

The account of the meeting continues : Many foreign colleagues were astonished. Baltzer von Born shook his head, others laughed. Some were outraged at Lugeon's absurd phantasies. The foremost Professor of Vienna, Uhlig, whose Carpathian work had been re-interpreted by Lugeon, suffered a nerve-shock. Then Heim was invited to reply. All listened tensely and were astonished by his agreement, all were influenced by the glory of the moment when Heim recognized the new truth, although in doing so the double-fold which he had defended for so many decades was demolished. Without hesitation or ill-humour, but with cheerfulness and stimulation, he agreed to the new teaching. It was one of the great moments in the history of science when the nappe theory was established before a gathering of distinguished geologists from all countries who were to carry their conviction to all ends of the earth.

Lugeon was the scientific, Heim the moral, victor.

The biography is attractively produced and illustrated with family photographs taken through the years, including one of three Heim generations, Arnold Heim the son being himself a geologist of distinction and versatility. Heim's skill in landscape drawing and geological field sketches is well demonstrated by a number of examples.

G. M. L.

PUBLICATIONS RECEIVED

Rocks and Mineral Deposits. By P. NIGGLI (translated by R. L. PARKER). xiii + 572 pp., 331 figs., 73 tables. W. H. Freeman and Company, San Francisco, 1954, price 102*s*.

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